Characteristics of Hawaiian Volcanoes
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Chapter 9

A Century of Studying Effusive Eruptions in Hawai‘i

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Abstract

The Hawaiian Volcano Observatory (HVO) was established as a natural laboratory to study volcanic processes. The most frequent form of volcanic activity in Hawai‘i is effusive; for this reason, a major contribution of the past century of research at HVO has been to describe and quantify lava flow emplacement processes. Lava flow research has taken many forms; first and foremost has been collection of basic observational data on active lava flows from both Mauna Loa and Kīlauea volcanoes that have occurred over the past hundred years. Both the types and quantities of observational data have changed with changing technology; for this reason, another important contribution of HVO to lava flow studies has been the application of new observational techniques. Also important has been a long-term effort to measure the physical properties (temperature, viscosity, crystallinity, and so on) of flowing lava. Field measurements of these properties have both motivated laboratory experiments and presaged the results of those experiments, particularly with respect to understanding the rheology of complex fluids. Finally, studies of the dynamics of lava flow emplacement have combined detailed field measurements with theoretical models to build a framework for the interpretation of lava flows in numerous other terrestrial, submarine, and planetary environments. Here, we attempt to review all these aspects of lava flow studies and place them into a coherent framework that we hope will motivate future research.

Introduction

Overview

Rivers of lava are an iconic image of Hawaiian volcanism. With the frequent eruptions of Mauna Loa throughout the 19th and early 20th centuries, and the persistent activity of Kīlauea from 1969 to 1974 and since 1983 (fig. 1), Hawai‘i has served as one of the most important natural laboratories in the world for studies of lava flows. Furthermore, the higher flow rates associated with most eruptions of Mauna Loa, and the longer duration of eruptions from Kīlauea, have provided important insight into a wide range of flow emplacement processes. In this chapter, we review observations of lava flow activity over the past century, outline many of the techniques developed in Hawai‘i to map and analyze lava flows, and discuss the physical conditions of flow emplacement. Finally, we examine some of the contributions of Hawaiian lava flow studies to the investigation of planetary volcanism.

As demonstrated by the global perspective of the “Volcano Letter” (compiled in Fiske and others, 1987; herein abbreviated as “VL” followed by the appropriate issue number) and other Hawaiian publications, research in Hawai‘i has not gone on in a vacuum; instead, it has clearly been influenced by academic researchers throughout the United

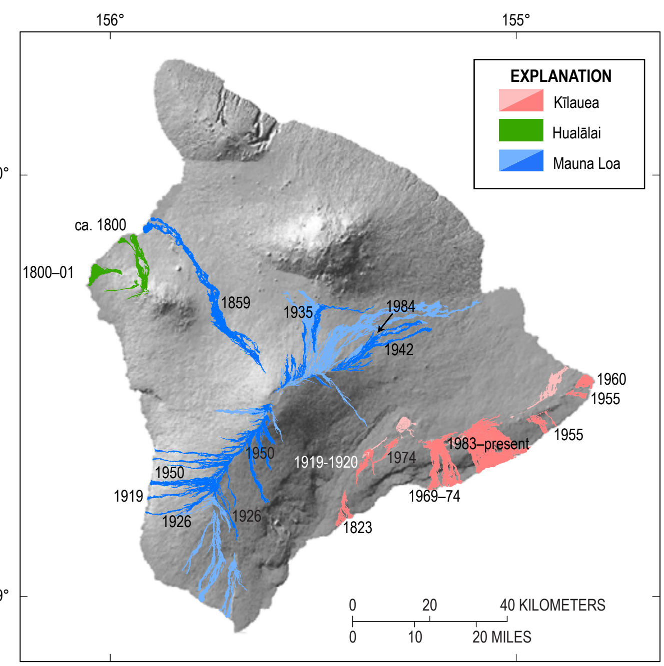

Figure 1. Shaded-relief map of Hawai‘i showing locations of all historical lava flows, color-coded according to source volcano. Flows mentioned specifically in text are labeled by date and shown in darker colors.

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States as well as by scientists studying volcanoes around the world. In fact, only the frequent lava flow activity at Mount Etna, Italy, is comparable to that in Hawai‘i in terms of its accessibility and impact on the study of effusive volcanism. Where appropriate, we note these global links, although here we focus primarily on the contributions to lava flow studies from research by, and in conjunction with, the Hawaiian Volcano Observatory (HVO).

The View from the 19th Century

The first written accounts of Hawaiian lava flows were by Ellis (1825), who traveled throughout Hawai‘i as a missionary in 1823. Another missionary—Titus Coan—provided detailed observations of most of the eruptions of Mauna Loa and Kīlauea between 1835 and 1882; many of these accounts were published in the American Journal of Science (see Wright and Takahashi, 1989, 1998; Barnard, 1990). In the 19th century, scientists, such as J.D. Dana, W.D. Alexander, C.E. Dutton, W.T. Brigham, and C.H. Hitchcock assembled their own observations and those of local observers (including Coan), so that by the end of that century the morphology of Hawaiian lava flows, the formation of lava channels and tubes, and the importance of cooling on flow evolution were all well described, albeit qualitatively. By the end of the century, a summary by Dana (1890) argued convincingly not only for Hawai‘i as a natural laboratory equivalent to that provided by Mounts Vesuvius and Etna in Italy, but also that the frequency and accessibility of Hawaiian lava flows “make them a peculiarly instructive field for the student of volcanic science, as well as an attractive one for the lover of the marvelous” (p. v). Dana proceeded to review key observations from 19th century Hawaiian studies, including (1) the eruption of (dense) olivine-phyric flows from high-elevation Mauna Loa vents, (2) the simultaneous activity of Mauna Loa and Kīlauea, (3) the “mobility” of liquid basalt, (4) the recognition of pit craters as characteristic of basaltic volcanoes, and (5) the activity of lava lakes. He also listed areas that required additional study, including the dynamics of lava streams (channels) and the formation of lava tubes, as well as the relation of Kīlauea to Mauna Loa and the driving force for eruptive activity. These broad topics identified by Dana have, indeed, framed volcanologic studies in Hawai‘i over the past 100 years.

In addition, 19th century volcanologists provided a foundation for more focused topics of research that continue to garner attention. One is the origin of the two primary Hawaiian lava flow types, pāhoehoe and ‘a‘ā. These words were adopted from the Hawaiian language and introduced into the volcanological lexicon to describe lava with smooth or broken (clinkery) surfaces, respectively (fig. 2). In his geologic summary of Hawai‘i, Dana (1849) noted that ‘a‘ā and pāhoehoe formed during “different phases in the volcanic action of one and the same period” (p. 162) with the only difference being “a variation in the rapidity of motion, or a renewal of movement from a cessation” (p. 163). Close observations of paired ‘a‘ā and pāhoehoe lavas from the 1859

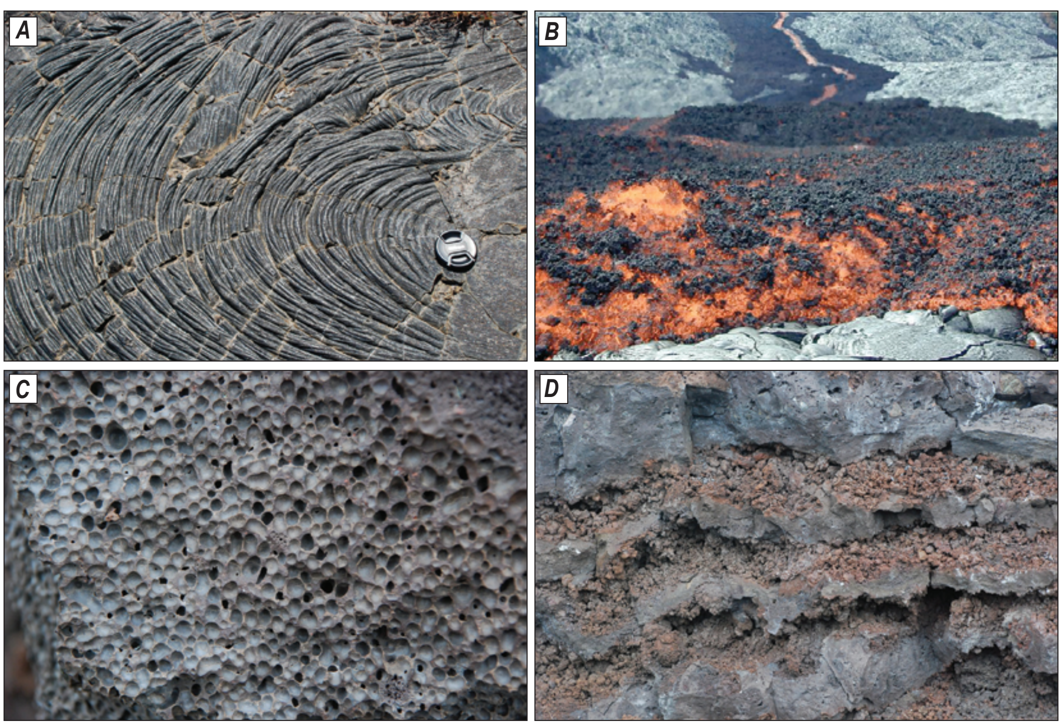

Figure 2. Photos of characteristic Hawaiian lava-flow morphologies. *A*, Ropy, degassed pāhoehoe flow surface. *B*, Active ‘a‘ā flow surface. *C*, Spongy pāhoehoe flow interior (individual vesicles are 2–5 mm across). *D*, Stacked thin ‘a‘ā channel overflows, showing characteristic dense cores (10–20 cm thick) and clinkery flow boundaries. All photographs by K.V. Cashman except fig. 2*B*, which is by S.A. Soule.

Mauna Loa eruption allowed the surveyor W.D. Alexander to extend this observation by inferring that the “mode of cooling” controlled the morphology, with thin, commonly near-vent pāhoehoe flows being in a “state of complete fusion,” while less fluid ‘a‘ā formed “grains like sugar” (Alexander, 1859). This metaphor has not only persisted through the HVO literature (for example, Jaggar, 1947; Macdonald, 1953) but also inspired the use of sugar crystallization (in fudge) as an analog for educational purposes (Rust and others, 2008). Alexander (1886) also recognized that the slope (flow rate) over which lava flowed was important in ‘a‘ā formation; thus, by the end of the 19th century, two primary controls on surface texture—crystallinity and flow rate—had already been identified.

Another long-lived topic of research begun in the 19th century was the quest to measure the physical properties of flowing lava (particularly temperature and viscosity). The spectacle of rapidly moving lava rivers of “irresistible impetuosity” captivated the earliest Western visitors to Hawai‘i (Ellis, 1825). Descriptions of lava as being “at a white heat and apparently as liquid as water” (Haskell, 1859) and of the “superior mobility” of Hawaiian lava, the relation of this mobility to lava temperature, and the role of mobility (viscosity) in processes of both eruption and flow emplacement (Dana, 1890) all illustrate the importance that early investigators placed on the relation between basic physical properties and the dynamics of lava flow emplacement.

A Summary of Effusive Activity from 1912 to 2012

The science of effusive eruptions has advanced in conjunction with both changes in the eruptive behavior of Kīlauea and Mauna Loa Volcanoes and the advent of new technologies. The first 5 decades of activity during the period 1912–2012 are recorded primarily in HVO documents, such as the “VL” and its early serial publications (compiled by Fiske and others, 1987; Bevens and others, 1988). More recent eruptions are thoroughly documented in HVO reports, U.S. Geological Survey Professional Papers, journal articles, and the HVO Web site (http://hvo.wr.usgs.gov/).

The First Five Decades

When HVO was founded in 1912, Mauna Loa was erupting at intervals of 3.5 years at the summit and 6 years on its flanks (VL 440), and so was considered the “grand theater for lava flows” (Hitchcock, 1911). This trend continued with important eruptions along Mauna Loa’s Southwest Rift Zone in 1919 and 1926 and Northeast Rift Zone in 1935 and 1942 (fig. 1). These eruptions were used by HVO scientists to (1) document the initiation and shutdown of activity at different vents and (2) monitor the rates and mechanisms of lava flow advance through steep and forested reaches, such as those of Hawai‘i’s southwest coast. Employment of U.S. military airplanes allowed scientists to obtain both real-time observations and aerial photographs of eruptive activity. Aerial capabilities prompted the first modern attempt at lava flow diversion in 1935, when bombs were dropped to disrupt robust lava tubes that were funneling lava flows toward Hilo (VL 431–432, 442, 445, 465, 506). Aerial observations in 1942 also permitted estimates of both early rates of flow advance (>6 mi in 5 hours [0.5 m/s]) and final flow volumes (2×10^9 ft^3 [5.7×10^7 m^3]; VL 476).

A summit eruption of Mauna Loa in 1949 preceded a massive eruption in June 1950 along the Southwest Rift Zone (VL 508, 509; Finch and Macdonald, 1953). Aerial and ground-based observations of the 1950 lava flow showed the source to be an en echelon fissure that stretched for nearly 20 km along Mauna Loa’s Southwest Rift Zone. The flow fed a wide area of anastamosing lava streams that plowed down the volcano’s forested southwest flank and poured into the ocean (fig. 3). Early flows covered the 24 km from the vent to the coastal highway in less than 3 hours (average flow advance rate, >2.2 m/s). HVO scientists were able to provide good estimates of both the area (35.6 mi^2 [91 km^2]) and volume (514×10^6 yd^3 on land, another 100×10^6 yd^3 in the ocean [470×10^6 m^3 total]) of the lava and to demonstrate that from one-half to two-thirds of the total volume was erupted during the first 36 hours of activity. These measurements yielded impressive initial (rather than average) effusion rates of ~4,700 yd^3/s (~3,600 m^3/s).

During the same period, activity at Kīlauea was restricted primarily to the caldera and Halema‘uma‘u Crater, which contained an active lava lake until 1924 (for example, Bevens and others, 1988). Effusive eruptions within the caldera were common but generally small in volume. One exception was

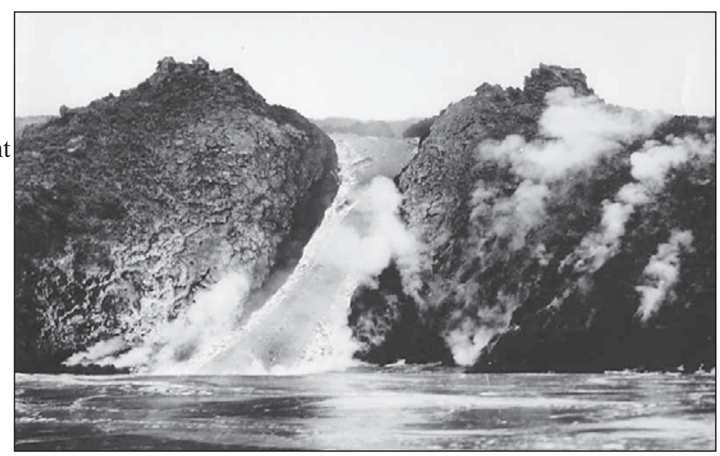

Figure 3. Photograph showing the Ka‘apuna flow, 1950 Mauna Loa eruption, entering the sea. Note modest amount of steam generated at ocean entry, as well as absence of explosions caused by lava-water interactions. Photograph by Transpacific Airways.

an eruption on the Southwest Rift Zone in 1919–20 (fig. 1), which produced paired ʻaʻā and pāhoehoe flows and the Mauna Iki shield (Rowland and Munro, 1993). This eruption is important in that it indicated a complex storage system beneath Halemaʻumaʻu with an intricate connection to Mauna Iki that evolved over time (see Bevens and others, 1988). Activity at Halemaʻumaʻu ended with a phreatic eruption in 1924; subsequently, Kīlauea Volcano outside of the summit was quiet until 1955, when a major flank eruption occurred on the East Rift Zone (Macdonald and Eaton, 1964; Moore, 1992).

The location of the 1955 eruption in the eastern Puna district (fig. 1), close to a seismic station installed in 1954, triggered a rapid response by HVO scientists. The proximity to populated areas helped HVO staff to locate the first flows, allowing them to be on site from the start of the eruption. From a perspective of geologic hazards, the 1955 eruption was the first to require extensive evacuations; as a result, it prompted the first modern attempts to construct barriers to deflect lava flows from critical areas (Macdonald, 1958). From a scientific perspective, this eruption provided scientists with their first opportunity to observe and photograph at close range the formation of a volcanic vent system, from the first opening of a fissure in the ground, through the appearance of lava, to the formation of cones and flows and, finally, the cessation of activity. Other opportunities provided by this eruption included observations of pit-crater formation and of active flow fronts at close range, which allowed study of the mechanics of flow movement and temperature measurements of both flow fronts and lava fountains (VL 529–530). Finally, this eruption provided an opportunity for direct comparisons between instrumental and field-based observations. Specifically, HVO scientists were able to correlate the amplitude of harmonic tremor with lava extrusion rate and record the progress of an eruption by monitoring changes in summit tilt.

The 1955 eruption was followed by a large summit eruption in Kilauea Iki pit crater in 1959 (see Helz and others, this volume, chap 6; Mangan and others, this volume, chap. 8) and, in 1960, by an eruption at Kapoho just downrift from the 1955 eruption site (fig. 1). Again, mitigation was attempted, with several barriers constructed in an effort to save homes and other places of historical interest (Macdonald, 1962). Lava ultimately covered an area of 10 km^2 (including all the barriers) with an estimated volume of 0.122 km^3 of new material.

1969–2012

Although numerous Hawaiian eruptions occurred during the period 1969–2012, those of Mauna Loa in 1984 and Kīlauea in 1969–74 (Mauna Ulu) and from 1983 to the present (Puʻu ʻŌʻō) are particularly well documented and have cemented HVO's reputation as a laboratory for studying basaltic volcanoes. These eruptions have provided new perspectives on old questions of lava flow emplacement, including measurements of the thermal efficiency of lava channels and tubes, the rheologic changes that accompany cooling and crystallization, the mechanisms of flow advance on both steep and shallow slopes, and the development of characteristic flow morphologies. Observers of these eruptions have also benefited from increasing access to eruption sites by helicopter, from the digital revolution (with its accompanying transformations in the acquisition, storage, and global transfer of a vast array of data), and from the application of numerous remote-sensing techniques.

Like many previous Mauna Loa eruptions, the 1984 eruption started at the summit (Mokuʻāweoweo) caldera on March 25. Within hours, eruptive activity migrated down the Northeast Rift Zone to establish a stable vent at 2,850-m elevation that fed the next 3 weeks of eruptive activity. The eruption, which produced 0.22 km^3 of lava and covered an area of 48 km^2 (fig. 4), illustrates characteristics that are typical of many "open channel" lava flows in Hawaiʻi. As eruptive vents migrated from east to west over time, they directed lava into different drainage basins, so that the flow direction shifted from east to northeast, toward Hilo. Each of the primary flows had a complex form, with numerous bifurcations (and some confluences), commonly around topographic barriers. Thus, Hawaiian lava flows are generally distributary, such that the total lava volume is divided between increasing numbers of flow lobes with distance from the vent.

From a geologic-hazards perspective, the advance of lava flows toward both the Kūlani prison and the city of Hilo caused some concern, although cooling ultimately arrested flow advance (Lockwood and others, 1987). From a scientific perspective, repeat observations at several places along the main lava channel provided unprecedented data on the flow of lava within the channel, as well as on the mechanisms of flow advance (Lipman and Banks, 1987); this unique dataset stimulated analysis of transport conditions through lava channels (Crisp and Baloga, 1994; Crisp and others, 1994) and continues to serve as a benchmark calibration for interpretations of older flows (Riker and others, 2009) and construction of flow models (for example, Harris and Rowland, 2001).

Two protracted eruptions on Kīlauea's East Rift Zone have provided similarly valuable datasets on the formation of compound pāhoehoe flow fields. The Mauna Ulu eruption (1969–74) produced 0.34 km^3 of lava that covered an area of 46 km^2, with the last 3 years of eruptive activity focused on the Mauna Ulu shield. Detailed observations of this eruption provided important new insight into shield formation and the characteristics of pāhoehoe lava (Swanson 1973), the behavior of shallow submarine lava flows (Moore and others, 1973), the pāhoehoe-to-aʻā transition (Peterson and Tilling, 1980), the formation of lava tubes (Peterson and others, 1994), and the dynamics of lava lakes (Tilling, 1987). The importance of this eruption has been somewhat eclipsed, however, by the unusually long Puʻu ʻŌʻō-Kupaianaha eruption farther down rift. This eruption, which began in early 1983 and is still ongoing as of September 2014, has provided a unique opportunity not only to study complex ʻaʻā and pāhoehoe flow fields but also to connect effusive activity to

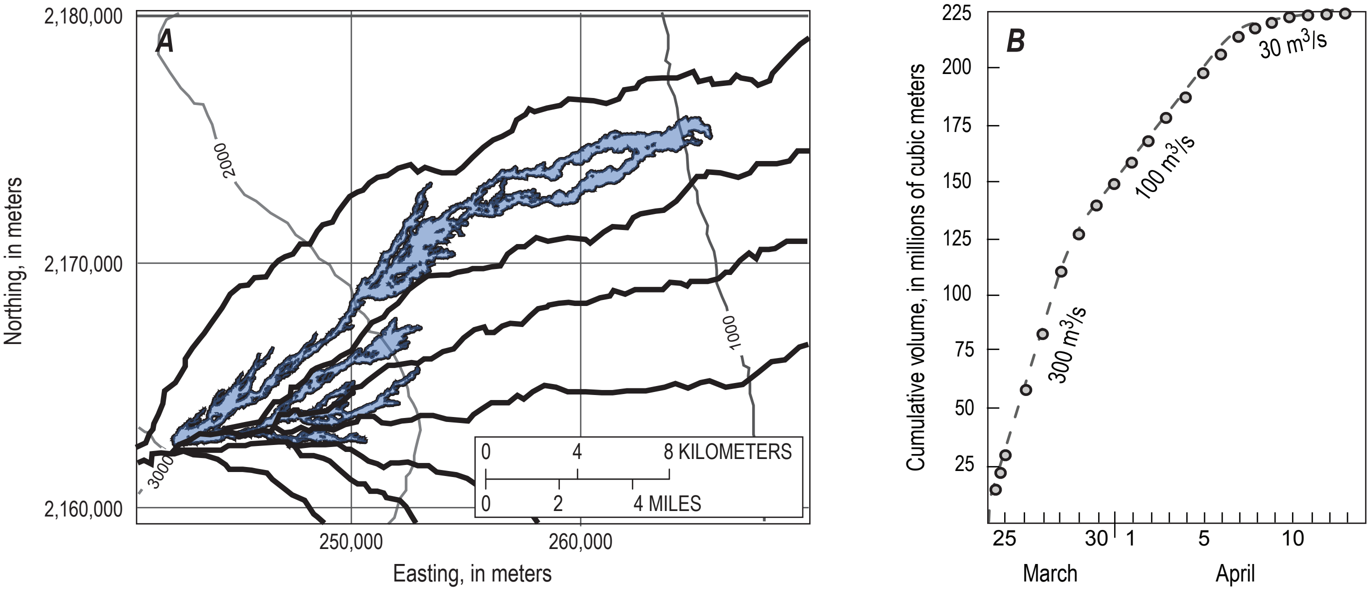

Figure 4. Lava flow from the 1984 Mauna Loa eruption (see fig. 1 for flow location). *A*, Sketch map showing extensive branching that accompanied flow emplacement. Heavy black lines denote “lava sheds,” or topographically defined drainage basins. Coordinates in UTM Zone 6, Old Hawaiian datum (modified from Kauahikaua and others, 1995). *B*, Cumulative erupted volume of 1984 flow versus time. On the basis of this plot, eruption can be divided into three phases: rapid effusion (300 m^3/s) for the first 5 days of activity, a protracted period of moderate effusion rates (100 m^3/s), and finally a waning stage (30 m^3/s; redrafted from Lipman and Banks, 1987).

the petrology, geophysics, and geochemistry of Kīlauea’s magmatic system (for example, Heliker and others, 2003; Poland and others, this volume, chap. 5). Below, we summarize observations from these eruptions on the basic characteristics of Hawaiian lava flows before addressing advances in more quantitative aspects of lava flow emplacement.

Basic Lava-Flow Characteristics

Hawaiian lava flows are commonly classified as either ‘a‘ā or pāhoehoe on the basis of their surface morphology (fig. 2), although also numerous transitional forms also exist. Flow type is not a function of composition but is controlled instead by processes related to the eruption and transport of lava (Macdonald, 1953). Here, we review observations on both flow morphology and flow geometry because they provide the framework for modern studies of lava properties, flow emplacement conditions, and hazard assessments.

Lava Morphology

Field geologists working in Hawai‘i have long been fascinated by the morphology of lava flow surfaces. Pāhoehoe, with its endless variety of surface forms, has spawned a proliferation of colorful descriptors such as sharkskin, toothpaste, rubbly, slabby, festooned, dendritic, shelly, blue glassy, pillow-like, and even entrail-like (for example, Stearns and Macdonald, 1946; Wentworth and Macdonald, 1953; Swanson, 1973; Rowland and Walker, 1987; Hon and others, 1994; Self and others, 1998; Kauahikaua and others, 2003). ‘A‘ā flows have not inspired a similar proliferation of names but also vary according to the size and shape of the surface clinkers (for example, Jones, 1943; Soule and others, 2004). Each morphologic variation reflects a set of intrinsic and extrinsic conditions that includes the specifics of the lava properties (particularly rheology) and external factors that control emplacement (such as volumetric flow rate, underlying slope, and topographic confinement, all of which contribute to the deformation conditions of the flowing lava). For this reason, investigation of the apparently simple question of crust morphology has led to a much deeper understanding of the thermal, rheologic, and dynamic evolution of lava flows.

Jaggar (VL 281) summarized the state of knowledge of ‘a‘ā and pāhoehoe derived from HVO studies covering the first 3 decades of the 20th century. By that time, he could state that “there is no essential difference chemically between aa and pahoehoe,” that “fountaining pahoehoe at the source of a flow may turn into aa clinkers within a half mile of the vent, and remain aa for the rest of its course down the mountain into the sea”, and that “when an observer stands on the bank of a golden, liquid torrent of lava flowing so rapidly as to make no crusts or skins, he cannot tell from the appearance of the liquid whether it will solidify as pahoehoe or aa.” He confirmed Alexander’s (1859) hypothesis that ‘a‘ā is more crystalline than pāhoehoe, and was able to extend this analysis to include the effects of stirring, with reference to the experimental data of Emerson (1926). Jaggar also recognized that the characteristics of pāhoehoe surface folds are determined by the “thickness of the flexible crust” at the time of deformation and that the vesicular crust of pāhoehoe “is an excellent insulator.” This work was later extended by application of folding analysis to the ductile layer (Fink and Fletcher, 1978), which can be used to determine the cooling and emplacement history of individual lava flows (for example, Gregg and others, 1998).

Jaggar's (VL 281) summary of the characteristics of ʻaʻā and pāhoehoe has stood the test of time. Subsequent experiments have reproduced Emerson's (1926) results, using different basaltic compositions and experimental conditions (for example, Kouchi and others, 1986; Sato, 1995); together, they show that shear (dynamic) crystallization is critical for ʻaʻā formation (see Rust and others, 2008). Macdonald (1953) assembled both a comprehensive description of the physical attributes of ʻaʻā and pāhoehoe flow morphologies and an extensive list of the conditions under which pāhoehoe lava could transform to ʻaʻā. He noted that flows change downslope from pāhoehoe to ʻaʻā but not the reverse (see Jurado-Chichay and Rowland, 1995, and Hon and others, 2003, for a more nuanced discussion of this point), that the distance lava travels from the vent before changing from pāhoehoe to ʻaʻā varies inversely with eruptive vigor (volumetric flow rate from the vent), and that pāhoehoe is more gas rich and hotter, and contains more quenched glass, than ʻaʻā.

Peterson and Tilling (1980) formalized these observations by defining the pāhoehoe-to-ʻaʻā transition as a threshold in the relation between shear-strain rate and apparent viscosity (fig. 5). Two views exist of this threshold. The perspective from observations of crystalline lava flows at Mount Etna is that it represents a failure envelope for flow crusts under conditions of continuous deformation (Kilburn, 1990, 1993). In Hawaiʻi, however, where lava emerges from the vent at near-liquidus temperatures, the importance of achieving a critical crystallinity is more apparent (for example, Rowland and Walker, 1990; Crisp and others, 1994; Cashman and others, 1999). Coupling of field and laboratory measurements with the results of analog experiments shows that the transition from pāhoehoe to ʻaʻā is determined primarily by a threshold value of apparent viscosity except at very low strain rates, where pāhoehoe morphologies can be maintained to higher viscosities (fig. 5). This assessment of the pāhoehoe-to-ʻaʻā transition presages recent rheologic studies that show the dependence of critical rheologic transitions on particle shape, volume fraction, size distribution, and strain rate (for example, Costa and others, 2009; Castruccio and others, 2010; Mueller and others, 2010, 2011; see subsection below entitled "Rheology").

Geometry of Flows and Flow Paths—An Observational History

The distribution of flow surface morphologies varies in both space and time, is directly linked to changing conditions of flow emplacement, and is determined by the type and geometry of lava transport systems. Flow surface mapping techniques have evolved in conjunction with changes in available technology, including first aerial, and then satellite, observational platforms that allow flow fields to be viewed, and analyzed, in their entirety.

The establishment of aerial monitoring in the 1930s permitted detailed observations of flow geometries and active flow surfaces. As a result, accounts of the 1935 eruption of Mauna Loa contain the first detailed descriptions of the intricate geometry of active lava channels and spatial changes in surface morphology from proximal braided pāhoehoe streams ("braided torrents of glowing liquid [that] were from 30 to 50 feet [10–15 m] wide, near their sources") to channelized ʻaʻā within about 1 mi (1.6 km) of the vents (VL 439). Observers also noted that the lava streams "became narrower further down the mountain," where "the rapidity of their forward motion became less," and that the flow surface showed a temporal progression as early ʻaʻā was covered by later pāhoehoe (VL 429).

The 1950 eruption of Mauna Loa afforded new opportunities to observe large channelized flows, although the steep forested flanks of Mauna Loa's Southwest Rift Zone limited most of these detailed observations to areas between the highway and the coast (fig. 3). Documentation of the 1950 lava flows included measurements of maximum channel flow rates over lava cascades (35 mi/h [>15 m/s]), standing waves 12 ft. (3.6 m) high below the cascades, and surges in flow advance, with flow rates (7–8 mi/h [3–3.5 m/s]) that exceeded those of normal channel flow (4–5 mi/h [~2 m/s]; Finch and Macdonald, 1953). Scientists noted the abundance of blocks transported through the channel, as well as their tendency to obstruct channels and create overflows. They also made numerous optical pyrometry measurements of flow temperature (see subsection below entitled "Temperature")

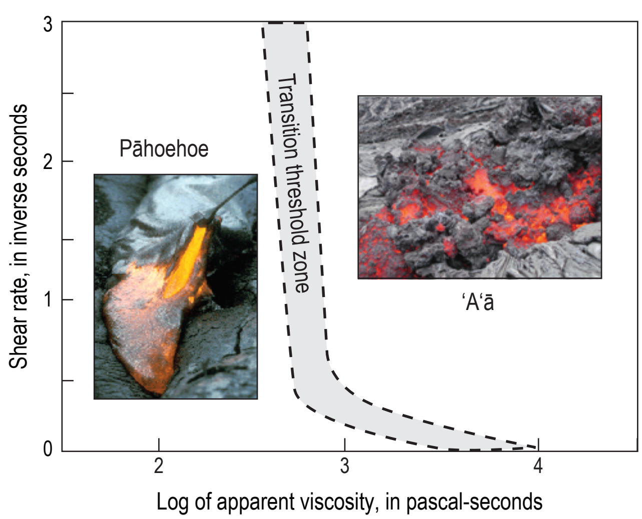

Figure 5. Plot of shear rate versus apparent viscosity fields for flow surface morphologies (pāhoehoe and ʻaʻā), calibrated from a combination of field observations (Cashman and others, 1999; Hon and others, 2003; Soule and others, 2004; Riker and others, 2009) and laboratory experiments (Soule and Cashman, 2005) and representing best approximation to date of pāhoehoe-to-ʻaʻā transition in Hawaiian lava flows.

and reported widespread methane generation from burning vegetation.

During the 1970s and 1980s, increased staffing at HVO and improved access to active flow fields (by trails, roads, and helicopters) brought new detail to flow field maps, including the time evolution of flow emplacement (for example, Lipman and Banks, 1987; Tilling and others, 1987; Wolfe and others, 1988; Heliker and others, 2001). These maps, which are now posted routinely on HVO's Web site, allow analysis of the relations between effusion rates, rates of lava-flow advance, and topographic confinement provided by earlier emplaced flows (for example, Kauahikaua and others, 2003). At the same time, helicopter support facilitated repeat observations in places along lava channels (Lipman and Banks, 1987) and lava tubes (for example, Mangan and others, 1995a; Kauahikaua and others, 1996; Heliker and others, 1998) that can be used to constrain the conditions of lava transport (see section below entitled "Dynamics of Lava Flow Emplacement").

The prolonged Pu‘u ‘Ō‘ō-Kūpaianaha eruption has produced a range of new flow mapping techniques. During the first decade of activity, maps were constructed directly on aerial photographs using helicopter surveillance and ground-based observations, coupled with post emplacement aerial photographs (for example, Wolfe and others, 1988; Mattox and others, 1993). The advent of hand-held Global Positioning System (GPS) units drastically changed mapping techniques by providing both accurate locations of flow features and digital data appropriate for use in geographic-information-system (GIS) mapping utilities. Conversion to digital mapping has improved the accuracy and efficiency of mapping efforts, especially as GPS maps of flow outlines can now be generated by helicopter surveys.

The 1990s also saw the application of remote-sensing techniques to near-real-time mapping of lava flows. Frequent (every 15 minutes) low-resolution (4 km/pixel) views of the lava flow field generated by the Geosynchronous Orbiting Environmental Satellite (GOES) allow tracking of hotspots related to surface activity (Harris and others, 2001). Advanced Very High Resolution Radiometer (AVHRR; 1 km), Landsat Thematic Mapper (TM; 30–120 m) and Moderate Resolution Imaging Spectroradiometer (MODIS) thermal images can be calibrated to yield estimates of the time evolution of lava effusion rates, which provide important input to predictive models (Flynn and others, 1994; Harris and others, 1998; Wright and others, 2001, 2002). Recent advances in flow mapping in both Hawai‘i and Italy include the use of airborne light detection and ranging (lidar) and satellite-based synthetic-aperture radar (SAR) for the generation of digital elevation models (DEMs; Rowland and others, 1999; Mazzarini and others, 2005), relative flow age determination (Mazzarini and others, 2007), and flow mapping (Zebker and others, 1996; Favalli and others, 2010; Dietterich and others, 2012; Cashman and others, 2013). Importantly, these data also provide new insights into flow field evolution by supplying detailed views of flow field construction.

Physical Properties of Flowing Lava

At the same time that HVO scientists were observing and mapping lava flows, they were also attempting to measure the physical properties of flowing lava and to link these properties to flow emplacement conditions. These properties include not only lava temperature but also changes in bubble and crystal content and their effect on lava rheology. Here we show how field-based observations and measurements of the physical properties of Hawaiian lavas have provided important data on the structure and rheology of silicate melts; have spurred laboratory and theoretical research on the relations between the temperature, rheological and material properties of mafic magma; and have provided key information on the rates and types of phase changes during lava transport. These data are critical for understanding the dynamics of lava flow emplacement.

Temperature

Early observers used the color of lava (for example, "white hot" versus "cherry red") to determine the relative temperature of different parts of individual lava flows. Use of a color scale to measure temperature was not unique to Hawai‘i; for example, Perret (VL 202) used a color-based scale to estimate a temperature of 1,200 °C for Etna lava from an eruption in 1908. E.S. Shepherd and F.A. Perret made the first direct temperature measurements of Hawaiian lava, using a platinum-rhodium thermocouple (Shepherd, 1912) to obtain a temperature of 1,000 °C for lava in the Halema‘uma‘u lava lake. Jaggar (1917, 1921) experimented with the use of Seger cones (used in firing pottery) to measure temperature-depth profiles within Halema‘uma‘u. The Seger cones, however, produced sufficiently confusing results that most workers relied on temperature measurements by optical pyrometer, which have yielded temperatures of 1,075–1,130 °C for Halema‘uma‘u fountains and flank vents, 1,120–1,190 °C for the unusually hot fountains accompanying the 1959 eruption of Kīlauea Iki, and 900–1,030 °C for channelized lava. Problems with these readings lie primarily in the difficulty in obtaining an unobstructed view of the fountain/flow interior and of knowing the appropriate correction for emissivity; as a result, it has commonly been assumed that optical pyrometry readings are 20–30 °C too low (Macdonald, 1963). In fact, in situ temperature (thermocouple) measurements at source vents of 1,140–1,147 °C (Mauna Loa 1984) and 1,110–1,150 °C (Pu‘u ‘Ō‘ō) are slightly higher than most optical pyrometry measurements, confirming Macdonald's (1963) suspicions. When conditions are optimal, however, temperature measurements made by thermocouple and two-color infrared pyrometer can agree to within 5 °C (Lipman and Banks, 1987); used together they allow documentation of the thermal history of lava fountains and flows in both space and time.

New experimental and analytical capabilities in the 1970s fueled a boom in the design and calibration of geothermometers that form the basis for modern petrologic investigations of magma storage conditions (for example, Blundy and Cashman, 2008; Putirka, 2008). Most useful for lava flow studies were glass geothermometers based on an observed linear relation between the temperature and MgO content of Kīlauea and Mauna Loa melts (Helz and Thornber, 1987; Montierth and others, 1995). These glass geothermometers have been used to examine spatiotemporal patterns in the temperatures of active and solidified flows (Cashman and others, 1994, 1999; Mangan and others, 1995b; Clague and others, 1999; Soule and others, 2004; Riker and others, 2009) and to constrain the thermal efficiency of lava tubes and channels (Helz and others, 1995, 2003, and this volume, chap 6; see subsection below entitled "Lava Tubes").

Short-term spatial and temporal variations in the temperature of lava flow surfaces are best captured by ground-based thermal imaging systems, such as forward-looking infrared (FLIR) (fig. 6). The potential of FLIR data is illustrated by a detailed study of small tube-fed pāhoehoe lobes formed at Kīlauea in August 2004 (Ball and others, 2008), where FLIR images were used to test models of stationary-flow cooling (for example, Keszthelyi, 1995a; Keszthelyi and Denlinger, 1996; Harris and others, 2005; Ball and Pinkerton, 2006) and to examine the effects of flow emplacement dynamics on heat loss. FLIR data document rapid initial cooling of pāhoehoe flow surfaces by radiative cooling and help explain the low apparent temperatures of flowing lava obtained by optical pyrometry. The FLIR data can also be used to test cooling models (for example, fig. 6*B*) and to link surface temperatures to changes in the material properties of lava. For example, Ball and others (2008) document pāhoehoe rope formation at $T_{surface}$ ~800 °C and development of crust strength at $T_{surface}$ ~700 °C, as manifested by the onset of flow inflation. These threshold temperatures are higher than those inferred from glass geospeedometry analysis, which suggest that ductile deformation can continue to temperatures as low as 627 °C as long as mechanical perturbations to the crust occur at sufficiently long time scales (1–10 s; Gottsmann and others, 2004). These temperatures are lower than those inferred from lava lake drill cores obtained by HVO scientists, and suggest a transition from ductile to brittle deformation at T~800 °C (Wright and Okamura, 1977).

Rheology

The first estimates of lava viscosity compared the velocity and depth of channelized lava with measurements of water flows (Becker, 1897; Palmer, 1927). The results of these early studies—that lava was only 10 to 60 times more viscous than water (that is, $1*10^{-2}$ to $6*10^{-2}$ Pa·s)—were orders of magnitude too low because the applied formulation assumed turbulent flow, which is appropriate for water but not for lava (Nichols, 1939; Wentworth and others, 1945). Subsequent application of laminar flow models derived apparent viscosity estimates of $3*10^3$ to $20*10^3$ poise (300–2,000 Pa·s) for Mauna Loa lavas and $2*10^3$ to $100*10^3$ poise (200–10,000 Pa·s) for Kīlauea lavas (Macdonald, 1963). Observational constraints on viscosity were limited, however, by both the accuracy of field measurements and assumptions of flow homogeneity. A critical but difficult measurement for lava viscosity estimates is that of flow depth, which cannot be measured directly (VL 480; Lipman and Banks, 1987). Also problematic is the complex thermal structure of an active lava flow, because assumptions of homogeneity ignore the formation of surface crusts. The end result is that field-based measurements yield only apparent (integrated) lava viscosities that are difficult to correlate with laboratory studies of homogeneous liquids or liquid-particle suspensions.

Detailed observations of the 1984 Mauna Loa lava flows confirmed that the apparent viscosity of vent lavas may be as low as 100–200 Pa·s. Apparent viscosity increases exponentially along the channel (fig. 7*A*) because of both internal

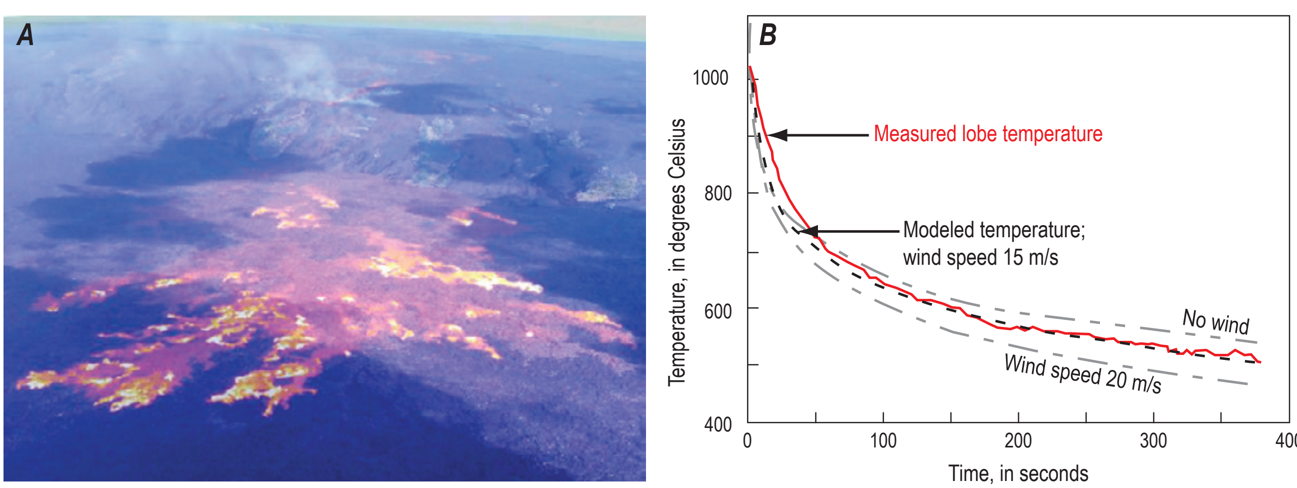

Figure 6. Temperature variation of lava-flow surface at Kīlauea. *A*, Composite of thermal image and color digital photograph. White and yellow areas, active pāhoehoe breakouts and red areas, inactive, but still warm, parts of flow surface (USGS photograph by Matt Patrick taken on April 26, 2012). *B*, Modeled and forward-looking-infrared lava flow surface temperatures versus time. Note rapid (radiative) surface cooling over first few minutes after initial lobe breakout (from Ball and others, 2008).

crystallization (fig. 7*B*) and crust formation (Moore, 1987; Crisp and others, 1994). Interestingly, the crystallinity data plotted in figure 7*B* show two different slopes (average crystallization rates). Earlier (higher) rates of crystallization over the first ~100 hours of effusive activity were apparently driven by uprift degassing (that is, the erupted lava was initially supercooled with respect to its temperature at atmospheric pressure because of the presence of dissolved volatiles). Degassing-induced crystallization continued for the duration of the eruption, but at a reduced rate as atmospheric equilibrium was approached (Crisp and others, 1994).

Experimental investigations of the rheology of Hawaiian basalts paralleled field-based measurements. In situ measurements by T.A. Jaggar, Jr., monitored the entry rate of lava into a metal cylinder (VL 357); in situ experiments were later conducted in Kīlauea’s lava lakes using a rotating shear viscometer (Shaw and others, 1968; see Mangan and others, this volume, chap. 8). In situ studies were complemented by laboratory studies on basaltic magma, although these were complicated by the need to employ different measurement techniques at high-temperature (low crystallinity and low viscosity) and low-temperature (high crystallinity and high viscosity) ends of the measurement spectrum (fig. 7*C* ; for example, Shaw, 1969; Murase and McBirney, 1973). Another type of in situ analysis involved settling of olivine phenocrysts through lava flows of different types (Rowland and Walker, 1987, 1988). This approach suggested a progressive increase in viscosity as lava morphology changes from smooth pāhoehoe (600–1,500 Pa·s) to rough pāhoehoe (6,000 Pa·s) to “toothpaste” lava (12,000 Pa·s). These values are about an order of magnitude higher than those calculated using measured melt temperatures and crystal contents (fig. 7*A* ; Riker and others, 2009). For this reason, estimates of the apparent viscosity controlling the threshold transition between pāhoehoe and ‘a‘ā shown in figure 5 are about an order of magnitude lower than that estimated on the basis of crystal-settling calculations.

Early workers also recognized that crystal-bearing magmas were not simple Newtonian fluids, and Bingham (1922) first suggested that his concept of yield-strength fluids (fig. 7*D*) might extend to lava. The idea that lava flows might have Bingham rheologies was extended by the burgeoning

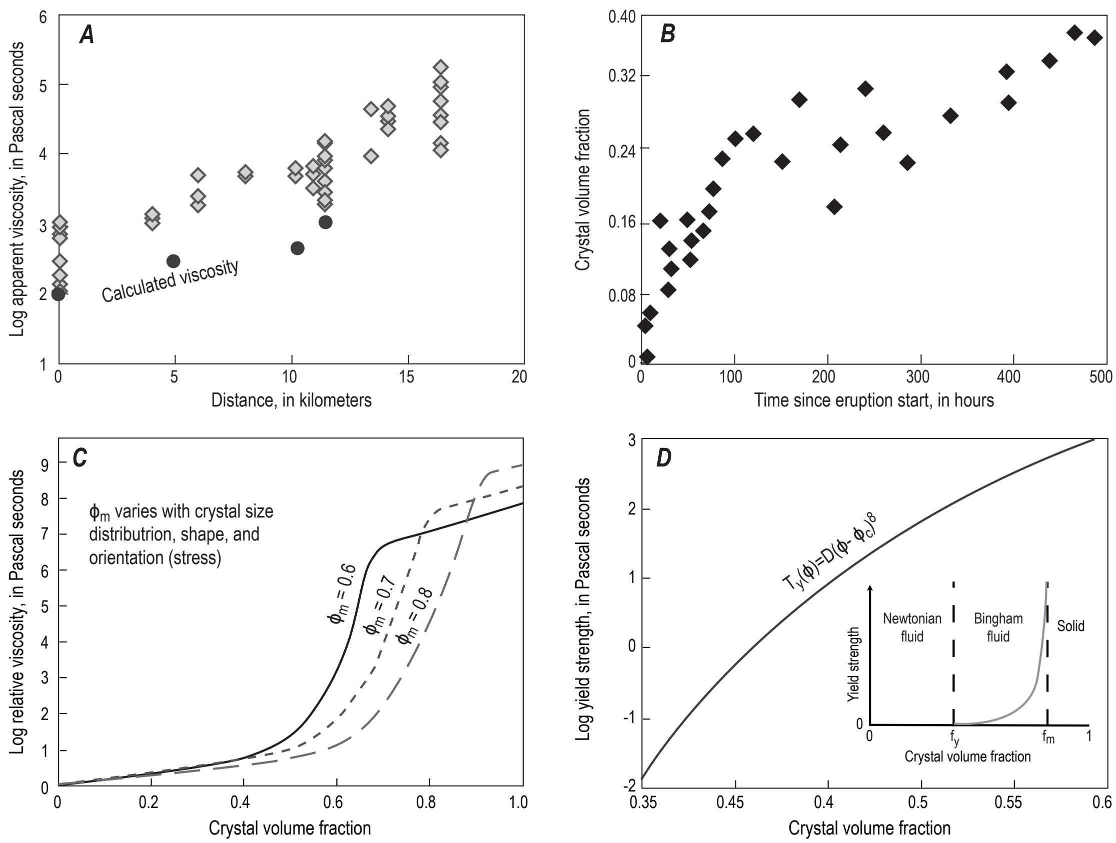

Figure 7. Lava flow rheology. *A*, Apparent viscosity (diamonds) versus distance along length of lava flow from 1984 Mauna Loa eruption, calculated using Jeffrey’s equation and data from Lipman and Banks (1987). Core lava viscosity (circles), calculated from measured glass temperatures and crystallinity (from Riker and others, 2009), is typically at least an order of magnitude lower than bulk apparent viscosity. *B*, Microphenocryst crystallinity of lava emerging from main vent of 1984 Mauna Loa eruption. Increasing crystallinity most likely reflects gas loss from magma during transport downrift (redrafted from Crisp and others, 1994). *C*, Relative viscosity (viscosity ratioed to pure-liquid values) versus critical, or maximum, crystallinity, Φ_m (redrafted from Costa and others, 2009). Note abrupt increase in viscosity close to critical crystallinity (“jamming point”). Absolute value of critical crystallinity depends on specifics of crystal population (distribution of crystal sizes, shapes, and orientations). *D*, Yield strength versus crystal content Φ beyond threshold crystallinity Φ_c for yield-strength onset; curve is a power-law function with a constant D=5×10^6 Pa (modified from Castruccio and others, 2010).

planetary-volcanological community in the 1970s. Hulme (1974) used analogue experiments to construct a model for the flow of a Bingham fluid on a slope. Using this model, he formulated a theoretical relation between channel formation and yield strength, where yield stress is defined as the minimum stress required for a homogeneous crystal-liquid suspension to flow. He tested his model using observations on lava flows from Mount Etna (Hulme, 1974) before using it to infer lava rheology on the Moon and Mars (Hulme, 1976; Hulme and Fielder, 1977).

Subsequently, Hulme's model has been applied to Hawaiian lava flows (for example, Fink and Zimbelman, 1986; Moore, 1987), where measured channel and levee dimensions suggest apparent yield strengths ranging from 0 to 5,000 Pa. Significantly, however, field estimates of yield strength in Hawaiian lava flows are complicated by many of the same problems that affect viscosity estimates (for example, Griffiths, 2000; Kerr and others, 2006). For this reason, field measurements based on lava flow properties should be considered apparent (or effective) values, and comparison with laboratory measurements should be made with caution.

In the laboratory, the onset of yield strength requires development of a "touching framework" of crystals (Kerr and Lister, 1991) that can bear stress, such that the crystal-melt suspension develops a viscoplastic rheology (Pinkerton and Sparks, 1978; Robertson and Kerr, 2012). At this point, the lava will cease to flow if the shear stress is sufficiently low, or may tear rather than deforming ductily under the imposed stress of continued downslope flow. The crystal volume fraction at which this transition occurs strongly depends on crystal shape and orientation (for example, Philpotts and others, 1998; Philpotts and Dickson, 2000; Hoover and others, 2001; Saar and others, 2001). Once the threshold crystallinity is achieved, yield strength increases as a power law function of crystal volume fraction (fig. 7*D* ; Castruccio and others, 2010).

Crystal shape also controls the maximum crystal volume fraction at which suspensions can continue to flow (generally designated the maximum packing fraction φ_m; Costa and others, 2009; Mueller and others, 2010, 2011). The deformation (shear) rate also affects suspension rheology by changing the spatial arrangement of crystals. For this reason, particle-melt suspensions are commonly modeled using a relation between applied stress (τ) and strain rate (γ) appropriate for Herschel-Buckley fluids (for example, Pinkerton and Norton, 1995). This treatment allows three fit parameters: the consistency, K (a measure of viscosity, determined by fitting φ_m for the suspension); the yield strength, τ_y; and the flow index, n, a measure of the extent to which the suspension is shear thinning or shear thickening. Also important, however, is the size distribution of crystals (Probstein and others, 1994; Castruccio and others, 2010; Cimarelli and others, 2011), which has not yet been fully incorporated into rheologic models (see review by Mewis and Wagner, 2009).

The presence of bubbles also affects the rheology of lava, although the magnitude of the effect is much less than that of crystals. Field observations suggest that bubble-rich lava can behave either more (Lipman and Banks, 1987) or less (Hon and others, 1994) fluid than its bubble-poor counterpart, depending on flow rate. These field observations are supported by laboratory experiments (Rust and Manga, 2002) and models (Pal, 2003; Llewellin and Manga, 2005) that demonstrate the relation between suspension viscosity and capillary number *Ca*, which is a measure of the extent to which the bubbles deform during flow. *Ca* is defined as $\mu V/\sigma$, where μ is the melt viscosity, V is the characteristic velocity, and σ is the interfacial tension between the gas and liquid phases. Thus, bubbles increase viscosity when strain rates are low (bubbles are undeformed) and decrease viscosity when strain rates are high (bubbles deform).

No models for magmatic systems account for the effects of both bubbles and crystals because of difficulties in modeling more than one suspended phase (for example, Tanner, 2009). Recent experiments on three-phase materials (both analogue and natural) indicate complex rheologies that may include both thixotrophic and viscoelastic behavior (for example, Bagdassarov and Pinkerton, 2004; James and others, 2004). Taken as a whole, the linkage between laboratory- and field-scale controls on rheology has been advanced by studies in Hawai‘i, but further research is needed to fully characterize the rheology of active lava flows.

Kinetics of Phase Change

Interest in the bulk properties (temperature and rheology) of lava led naturally to an interest in the bubbles and crystals present within the melt. Bubbles form in response to depressurization and, once formed, can move through fluid lava; for this reason the bubble content of lava can be used to monitor gas loss during eruption and emplacement. Crystal formation is sensitive to the rate and extent of lava cooling, as well as to stirring (for example, Emerson, 1926); thus, changes in the crystal population can be used to monitor the thermal and dynamical evolution of flowing lava.

Bubbles

Bubbles (or their frozen equivalents, vesicles) are nearly ubiquitous in the products of volcanic eruptions. Since the initial volatile composition of Hawaiian magma does not vary substantially, the bubble population in eruptive products is determined primarily by the vesiculation history (controlled by the decompression path) and subsequent patterns of bubble escape. In general, pyroclasts formed in Hawaiian lava fountains contain more and smaller bubbles than lava flows with the same total vesicularity generated by purely effusive activity (for example, Cashman and others, 1994; Mangan and Cashman, 1996; Stovall and others, 2011); this variation is inferred to reflect higher rates of decompression during fountaining eruptions (as measured by observed variations in mass eruption rate, fig. 8*A*).

Early workers recognized five different styles of vesiculation, two of which characterized lava fountain eruptions (see Mangan and others, this volume, chap. 8) and three possibly related to different lava emplacement styles (Hitchcock, 1911); these categories include (1) “ordinary lava streams,” with ≤60 volume percent elongate vesicles; (2) “spherically vesiculated lava,” with 30–60 volume percent bubbles in the upper parts of the flow; and (3) “the scum of the lava, which is often troublesome because one breaks through it in walking,” with 65–75 volume percent bubbles. The first vesiculation style, which is considered to be diagnostic of ‘a‘ā flows (Macdonald, 1953; fig. 2*D*), reflects both the high crystallinity (viscosity) and large shear strains applied to lava that has been transported through open channels (for example, Cashman and others, 1999; Soule and others 2004; Riker and others, 2009). The second style, which is also common, is diagnostic of inflated pāhoehoe flows (for example, Hon and others, 1994; Katz and Cashman, 2003). In these flows, the vesicular upper crust forms while the flow is actively inflating. In contrast, thin (rapidly quenched) pāhoehoe lobes may show a uniform distribution of spherical vesicles throughout (spongy pāhoehoe; Walker, 1989; fig. 2*C*). The third style, lava “scum,” is termed shelly pāhoehoe (Jones, 1943; Wentworth and Macdonald, 1953) and is common near eruptive vents (for example, VL 502). Shelly pāhoehoe forms by gas accumulation beneath thin, rapidly formed crust (Swanson, 1973).

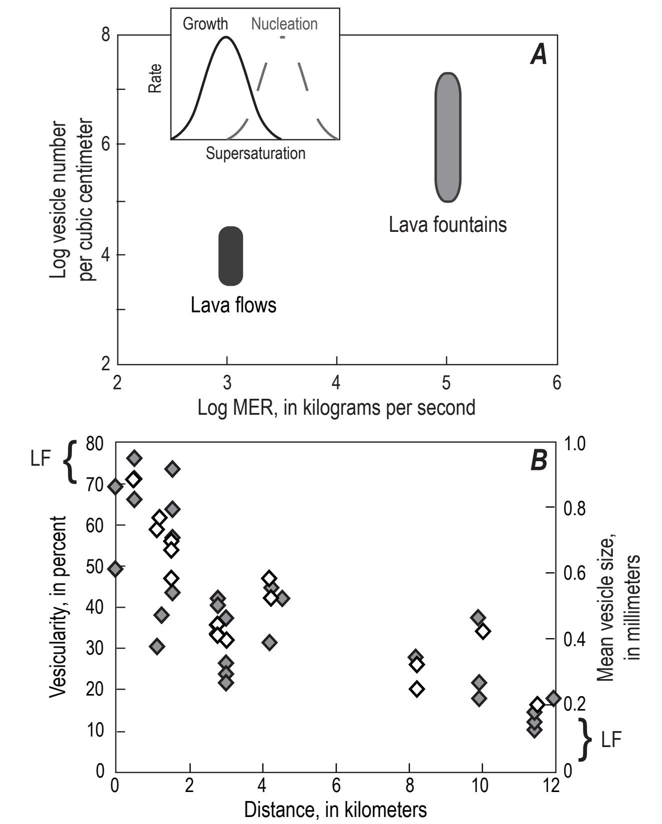

Figure 8. Vesicularity of Hawaiian lava. *A*, Plot showing bubble number density in lava-flow and lava-fountain samples. Higher bubble number densities record higher rates of bubble nucleation, which are associated with more rapid magma ascent (higher decompression rates), as inferred from estimated mass eruption rate (MER); data from Cashman and others (1994), Mangan and Cashman (1996), and Stovall and others (2010). *B*, Plot of lava vesicularity (gray diamonds) and mean bubble size (open diamonds) versus distance along two Kīlauea lava tubes. Steady decrease in both parameters reflects escape of gas bubbles from lava free surface. Redrafted from Cashman and others (1994). Brackets, ranges of vesicularity and mean bubble size reported by Mangan and Cashman (1996) and Stovall and others (2010) for lava-fountain (LF) samples.

Later workers have added to the lexicon of vesicle textures by identifying “flow pumice” and “P-type” pāhoehoe. Flow pumice is described as a tan, highly vesicular (~75 volume percent vesicles) glass skin on short flows adjacent to fissures. For example, pumice-surfaced pāhoehoe that formed during the 1942 eruption of Mauna Loa was “restricted to gushes of lava which traveled less than a quarter of a mile from their vents” and inferred to have formed by rapid (“nearly explosive”) vesiculation during emplacement (VL 502). Near-vent flow pumice produced during the 1859 eruption of Mauna Loa approaches reticulite in both vesicularity and structure, supporting this interpretation. P-type pāhoehoe is characterized by the presence of pipe vesicles in the lower parts of the flow (Wilmoth and Walker, 1993) and commonly occurs as breakouts from hummocky tumuli, where lava stagnates and partially degasses (Swanson, 1973). Pipe-vesicle-bearing flows tend to be dense, with “blue glassy surfaces,” probably a consequence of bubble loss during temporary lava residence within hummocky flows before final emplacement (Hon and others, 1994). The pipe vesicles grow inward from the cooling front as “cold fingers” (for example, Philpotts and Lewis, 1989), and although they are commonly seen at the flow base, they may also form radially around flow margins. This observation suggests that the use of pipe vesicles as flow-direction or paleoslope indicators (Waters, 1960; Walker, 1987) must be done with caution (Peterson and Hawkins, 1972; Swanson, 1972).

To date, field measurements also provide most of the constraints on the vesiculation kinetics of basaltic magmas (for example, Mangan and others, 1993, this volume, chap. 8, Mangan and Cashman, 1996) because the rapidity of bubble formation makes laboratory experiments challenging (see Murase and McBirney, 1973; Pichavant and others, 2013). Direct measurements of bubble populations in flowing lava have been made along both the 1984 Mauna Loa channelized flow (Lipman and Banks, 1987) and robust lava tubes that formed during the Pu‘u ‘Ō‘ō eruption (Mangan and others, 1993; Cashman and others, 1994). Both datasets show a downflow decrease in bubble content with distance from the vent that is consistent with observations of gas escape from flow surfaces (for example, fig. 8*B*). In lava tubes, vesicularity decrease is accomplished primarily by loss of large bubbles, as observed in lava tubes, described as “an orange-hot cavity with a golden river sweeping by underneath, little bubbles continually breaking the surface of the glowing stream, and adding gas to the evenly brilliant walls” (VL 345). Bubbles frozen into the growing upper crust of inflated lava flows can also record pressure changes within the tube during flow emplacement (Cashman and Kauahikaua, 1997; see subsection below entitled “Lava Tube Formation”).

Crystals

As reviewed above, Hawaiian lavas are commonly erupted at near-liquidus temperatures; for this reason, they can crystallize extensively, particularly during flow through open channels. Early 20th century crystallization studies were aimed at defining conditions of ‘a‘ā and pāhoehoe formation (for example, Emerson, 1926). Then, five decades later, a renewed interest in crystallization kinetics accompanied the advent of the semiconductor industry (for example, Kirkpatrick, 1981) and the collection of lunar samples (for example, Dowty, 1980; Basaltic Volcanism Study Project, 1981). Hawaiian lava lake samples, in particular, provided a well-controlled natural laboratory for these studies (see Helz and others, this volume, chap. 6; Mangan and others, this volume, chap. 8).

More recently, protracted lava flow eruptions have allowed relatively easy access to lava flows. Analysis of quenched samples from these flows can be used to link crystallization conditions directly to cooling rates (for example, Crisp and others, 1994; Cashman and others, 1999; Soule and others, 2004; Riker and others, 2009). Typical Hawaiian lava is erupted at temperatures of 1,150–1,170 °C, when the melt is multiply saturated with plagioclase and pyroxene. Under these conditions, measured cooling rates of ~0.005 °C/s along near-vent open channels drive crystallization at a rate of ~0.005 to 0.01 volume percent per second (18–36 volume percent per hour). At these high cooling rates, crystallization occurs primarily by nucleation of new crystals rather than by growth of existing crystals (fig. 9*A*). The dominance of crystal nucleation is illustrated by a steady increase in the measured number of both plagioclase and pyroxene crystals with increasing crystal volume fraction (fig. 9*B*), as well as by patterns of crystal-size distributions (CSDs). Samples collected along a single lava channel on the same day show parallel CSD trends (fig. 9*C*). Here, the slope of the line provides a measure of dominant crystal size, and the area under the line is the total number of crystals; therefore, parallel trends reflect addition of crystals of the same (small) dominant size (for example, Cashman and others, 1999).

In contrast, slow cooling accompanying lava transport through well-insulated lava tubes or solidification of stable lava lakes promotes crystal growth over crystal nucleation. Growth-dominated crystallization is manifested by the maintenance of constant crystal numbers with transport distance (fig. 9*A*) and total crystallinity (fig. 9*B*). CSDs also show patterns characteristic of growth-dominated crystallization, such that CSDs pivot around a point (fig. 9*D*) rather than showing the parallel trends plotted in figure 9*C*. Fanning CSDs record increases in dominant crystal size (inversely proportional to slope) with increasing volume fraction at either constant, or even decreasing, total crystal numbers (Cashman and Marsh, 1988).

Quenched samples can also provide insight into the relation between crystallization and flow-surface morphologies, which, in turn, place constraints on flow rheology (for example, Cashman and others, 1999; Soule and others, 2004; Riker and others, 2009). Such studies suggest that smooth pāhoehoe surfaces can be maintained to groundmass crystallinities of <~20 percent if shear rates are sufficiently low. Flow surfaces can start to develop transitional (rough) surface characteristics at groundmass crystallinities as low as ~15 percent if shear rates are higher; groundmass crystallinities > 35 percent typically have fully formed ‘a‘ā textures regardless of shear rate (fig. 10*A*). These observations can be explained by (1) a shear-rate-dependent onset of yield strength at 10–20 volume percent crystals and (2) a drastic increase in viscosity at ~35 volume percent crystals, suggesting a relatively low value of φ_{max} in Hawaiian lavas. This low value of critical crystallinity (compare with fig. 7*C*) is consistent with the high anisotropy of groundmass plagioclase crystals. Support for this interpretation comes from observations of lava flows that maintain pāhoehoe surfaces to moderately high crystal contents (~50 volume percent). These flows (such as the picrites common on Kīlauea and Mauna Loa’s rift zones and the “cicirara” [chickpea] lavas of 17th century Mount Etna eruptions) are dominated by near-isotropic phenocrysts, which should have a “jamming” point in excess of 40–50 volume percent.

Together, the studies outlined above show that (1) channelized Hawaiian lavas crystallize rapidly during early stages of transport; (2) rapid crystallization produces numerous small crystals that create an abrupt increase in the effective viscosity of the lava; and (3) subsequent deformation of the magma occurs by tearing rather than ductile flow, thereby creating the rough surface that is characteristic of ‘a‘ā. Solidified ‘a‘ā flows are finely crystalline throughout (fig. 10*B*), testifying to the efficiency of stirring in these flows (Griffiths and others, 2003; Cashman and others, 2006). In contrast, tube-fed surface flows commonly have glassy pāhoehoe surfaces, a consequence of minimal syntransport cooling and crystallization. Postemplacement crystallization textures mirror cooling rates, so that glassy pāhoehoe surfaces rapidly transform to first finely and then coarsely crystalline interiors (fig. 10*B*; Oze and Winter, 2005) with declining rates of cooling. Contrasting crystallization textures preserved in channelized and tube-fed lava flows thus provide critical information on both syneruption and postemplacement cooling; the resulting textural contrasts are particularly useful for interpretation of drill-core samples that lack an areal context (Katz and Cashman, 2003).

Dynamics of Lava-Flow Emplacement

Understanding where lava flows are likely to go, how far a given flow will travel, and how quickly it will advance are questions that are critical to the assessment of lava-flow

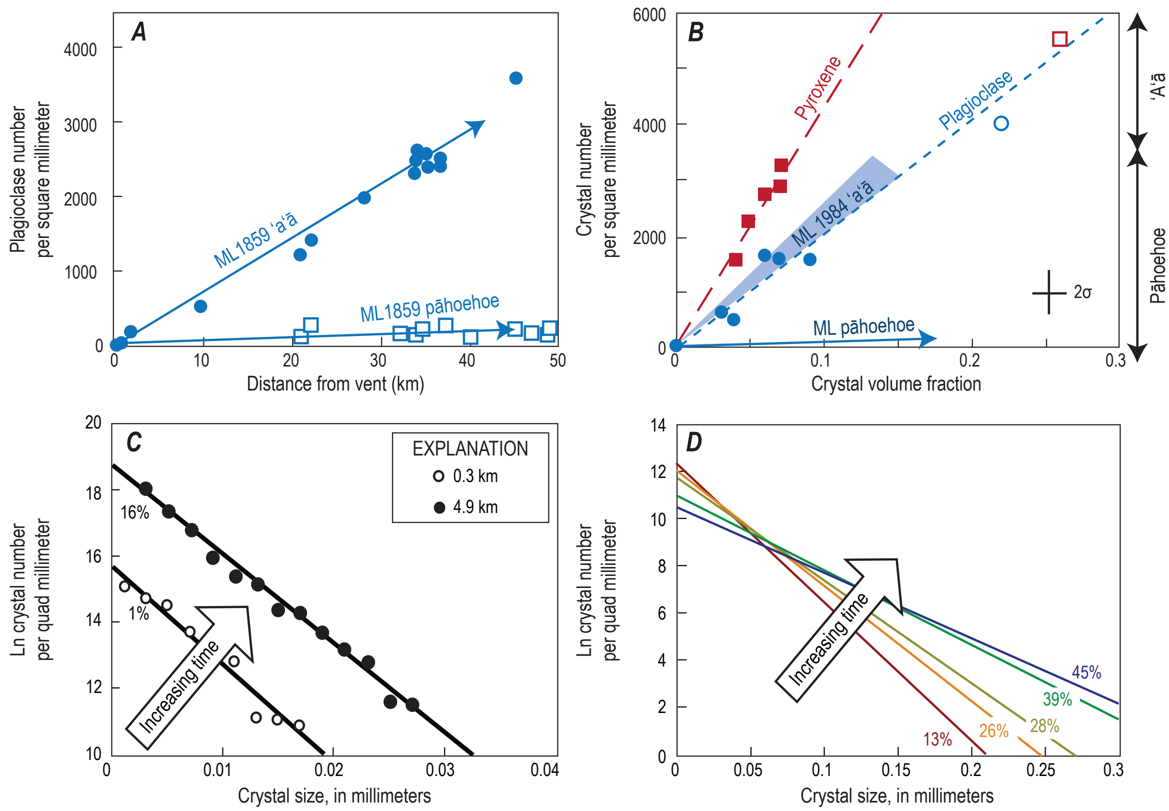

Figure 9. Crystal textures in Hawaiian lava flows. *A*, Plot of plagioclase microlite number density versus distance along paired ‘a‘ā (blue circles) and pāhoehoe (open squares) flows from 1859 Mauna Loa (ML) eruption. Two contrasting trends represent nucleation (‘a‘ā) and growth (pāhoehoe)-dominated crystallization (redrafted from Riker and others, 2009). *B*, Plot showing microlite number densities of plagioclase (blue dots and circles, lines) and pyroxene (red squares, line) versus changes in crystallinity. Data from small ‘a‘ā flow from the 1997 Kīlauea eruption. Closed symbols, samples with pāhoehoe flow surfaces; open symbols, samples with transitional flow surfaces (Cashman and others, 1999). Lines and shaded field are data from paired ‘a‘ā and pāhoehoe flows from the 1984 Mauna Loa eruption (from Riker and others, 2009). *C*, Plot of plagioclase crystal size distributions (CSDs; for example, Cashman and Marsh, 1988) from Episode 16 of the Pu‘u ‘Ō‘ō eruption, collected at different sites along main lava channel. Explanation indicates distance from vent. Parallel trends are another reflection of nucleation-dominated crystallization in channelized ‘a‘ā flows. *D*, Plagioclase CSDs of samples collected through crystallization interval of Makaopuhi lava lake (reanalyzed samples from Cashman and Marsh, 1988). Plagioclase crystallinity is labeled for each line; fanning CSDs show growth-dominated crystallization trends.

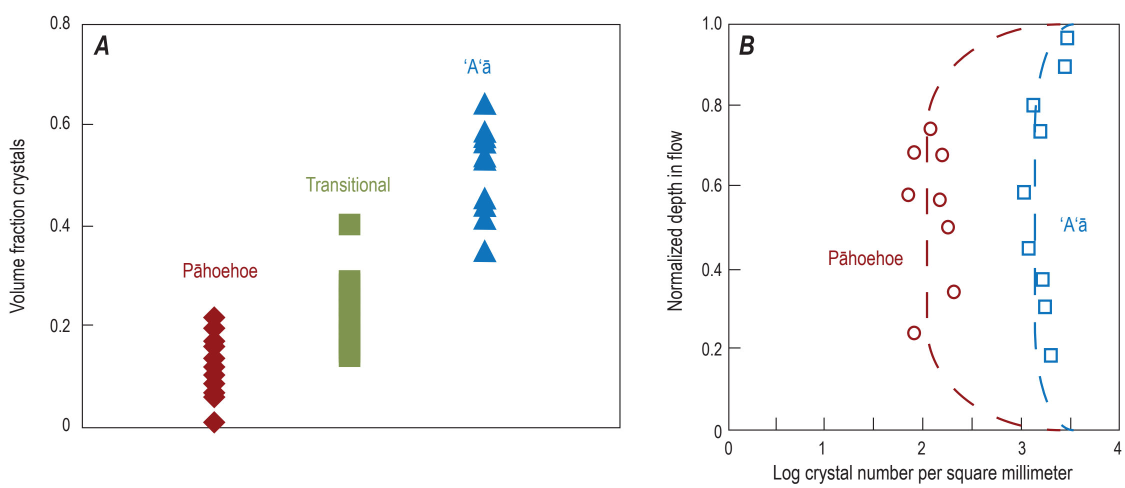

Figure 10. Properties of selected lava samples collected along individual open-channel flows during episodes 1–20 of the Pu‘u ‘Ō‘ō eruption (from Folley, 1999 and Cashman and others, 1999). *A*, Plot of total groundmass crystallinity of samples with different surface textures. *B*, Plot of crystal number density versus normalized depth through typical pāhoehoe (circles) and ‘a‘ā (squares) flows; dashed lines show approximate trends of data points. In solidified flows, crystal number density varies inversely with crystal size (redrafted from Katz and Cashman, 2003).

hazard. The answers to these questions differ for the short, high-effusion-rate channelized lava flows from Mauna Loa and the long-lived pāhoehoe flow fields that characterize current (and previous; for example, Clague and others, 1999) eruptive activity at Kīlauea. In this section, we review the physical controls on both styles of flow emplacement and illustrate some of the ways in which flow models are being used for hazards assessment.

Open-Channel Flows

When lava effusion rates are moderate to high, incandescent "rivers" of lava flow downhill, confined by lateral levees of lava rubble. In Hawai'i, these channelized flows are typically emplaced over time scales of hours (for example, 1974 Kīlauea; Lockwood and others, 1999) to weeks (for example, 1984 Mauna Loa; Lipman and Banks, 1987). The first detailed measurements on open-channel flows (including channel velocities, flow volume, and effusion rate) were made during the 1919 and 1926 Mauna Loa eruptions (VL 480). As described previously, these eruptions also saw the advent of aerial observations, which permitted not only descriptions of the plan-form geometry of braided lava streams but also details of flow surfaces. With the increase in real-time observations came questions about the construction of, and flow through, lava channels, as well as about the evolution of lava flux and channel geometries over time; these processes must be understood for predictions of flow length and aerial coverage. Although many parameters important for characterizing channel-fed flow are now routinely measured, some key measurements, such as the depth of lava flowing in channels, are still poorly constrained.

Morphology of Channelized Flows

Channelized flows commonly initiate from fissure vents, from which they travel as broad lava sheets that focus into lava channels after a flow of tens to hundreds of meters. Early proximal flows form anastomosing channels of hot fluid lava with thin surface crusts and small marginal levees (fig. 11*A*). Flow away from the source vents causes rapid cooling, crust formation and continued levee construction. Detailed observations of the 1984 Mauna Loa lava flow showed that stable channels form from a zone of dispersed flow at the propagating flow front through a transitional zone (Lipman and Banks, 1987). The stable channel may evolve over time as the flow focuses by inward solidification and as channel surges and (or) blockages create overflows (fig. 11*B*).

Open-channel flows are distributary, and the mass of flowing lava decreases along the channel from the vent to the flow front. This process is illustrated by measurements made along the 1984 Mauna Loa channel on a single day (April 4, 1987; fig. 11*C*), which show that bulk (lava plus bubbles) volumetric flow rate through the channel dropped by a factor of 5 over a distance of 15 to 20 km. This decrease was caused, in part, by loss of bubbles along the flow; however, even when corrected for changes in vesicularity, the volumetric flux decreased by a factor of 3. This loss of volume shows the extent to which lava channels are prone to mass loss by both overflows and storage within stagnant or near-stagnant marginal parts of the channel system (fig. 11*B*).

Overflows form when the channel is constricted or blocked by surface crusts or rafted accretionary lava balls, or when temporary increases in flux exceed the channel's carrying capacity. Overflows may develop into new flow branches if sufficiently sustained; for this reason, channel systems are typically distributary in plan form (fig. 4*A*). Individual channels generally widen as slopes decrease, and

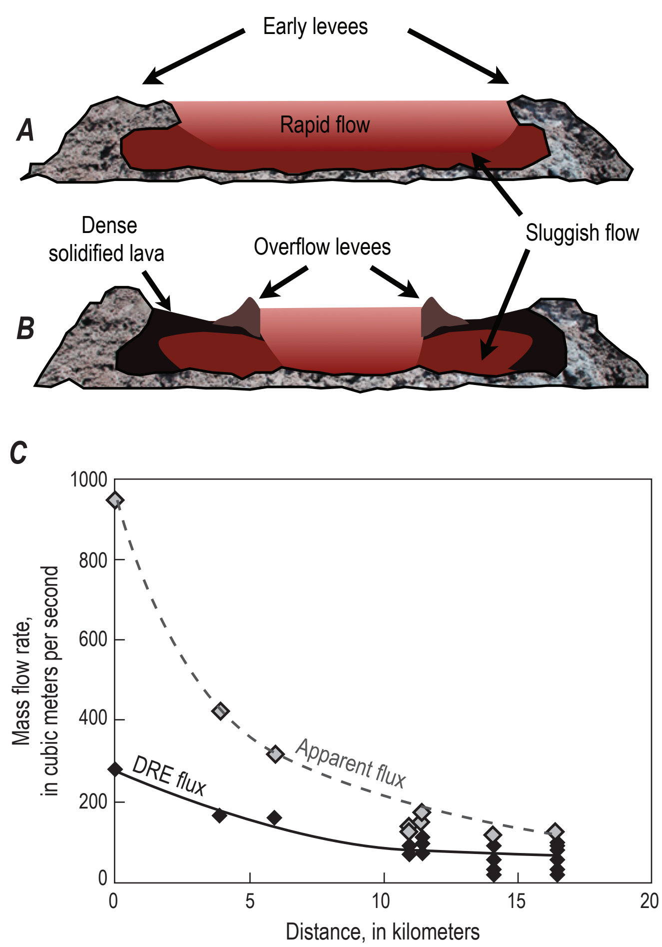

Figure 11. 'A'ā channel formation. *A*,*B*, Typical cross sections through early and later flow stages. Early flow (*A*) is characterized by outer levees surrounding inner fluid lava, which is separated from levees by a zone of sluggish lava flow. Later channel (*B*) has evolved to include solidified lava separating flowing core lava from levee; an inner channel has also formed, separated from original channel by overflow levees (redrafted from Lipman and Banks, 1987). *C*, Volumetric flux versus distance along lava channel from Mauna Loa eruption on April 4, 1984. Apparent flux is calculated from measured flow width, depth, and velocity; dense rock equivalent (DRE) flux is corrected for variations in lava vesicularity. Data from Lipman and Banks (1987).

narrow as effusion rates decline (Kerr and others, 2006). Posteruption surveys of the 1984 Mauna Loa flow showed channel floors to be close to preflow surfaces, suggesting that initial flow material is eroded as channels mature, but that erosion does not ordinarily extend into older, colder rock. An exception may be the c. 1800 Kaʻūpūlehu lava flow from Hualālai volcano (fig. 1), where channels show evidence of mechanical erosion during transport of large dunite xenoliths (Kauahikaua and others, 2002).

Cooling Along Lava Channels

Radiative heat loss from exposed lava surfaces is responsible for the bulk of the heat loss from most lava channels. Application of glass geothermometry shows that initial cooling rates may be as high as 0.01–0.03 °C/s where surface crusts are very thin and eruptive temperatures high (Cashman and others, 1999; Riker and others, 2009). Maintenance of exposed fluid lava along the channel margins thus requires efficient convection from the flow interior (Griffiths and others, 2003). Cooling rates drop to ~0.005 °C/s along medial to distal ʻaʻā channels (Crisp and others, 1994; Cashman and others, 1999), reflecting both the formation of an insulating ʻaʻā crust and buffering of lava temperatures by latent heat of crystallization.

Controls on Flow Length and Advance Rate

The length of a simple channelized lava flow will increase with lava effusion rate if the maximum flow length is controlled by cooling, if cooling rates are constant, and if effusion rate controls flow velocity (for example, Walker and others, 1973; Pinkerton and Wilson, 1994; Harris and Rowland, 2009). Alternatively, flow length may be limited by eruptive volume (flow duration), such that the flow does not reach its cooling-limited extent. In general, short duration flows tend to be limited in length by lava supply (“volume limited”), whereas the lengths of long-lived flows are limited by cooling (“cooling limited”).

Most Hawaiian lava flows are not simple in this sense, however, as illustrated by the lack of correlation between channelized-flow lengths and effusion rates, even for flows of similar duration (fig. 12). There are several possible explanations for this. First, the rheology of Hawaiian lavas at the vent may vary widely with bulk composition, temperature, crystallinity, and bubble content (Riker and others, 2009). Additionally, the high fluidity of Hawaiian lavas makes them susceptible to topographic confinement that will promote flow lengthening (Soule and others, 2004), whereas channel bifurcations caused by topographic obstacles create multiple parallel channels that can limit individual flow-lobe lengths (Lockwood and others, 1987). Finally, flow fields generally widen rather than lengthen when magma supply is unsteady and flows are emplaced as discrete events (Guest and others, 1987; Wolfe and others, 1988; Kilburn and Lopes, 1991; Heliker and others, 1998).

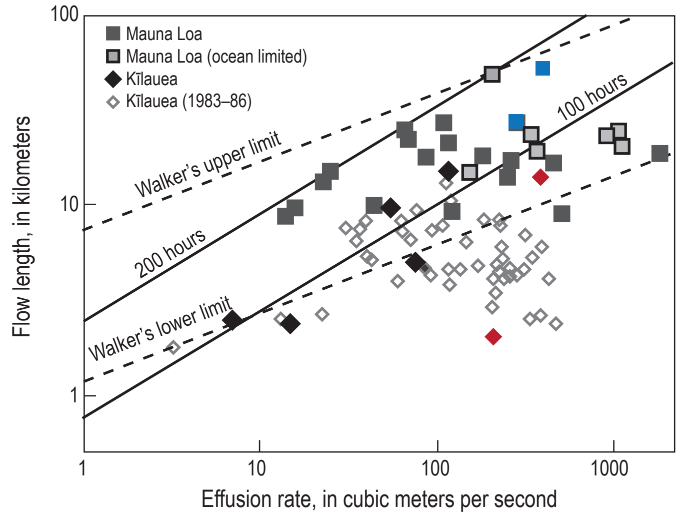

Figure 12. Plot of the lengths of channelized Hawaiian lava flows as a function of average volumetric flow (effusion) rate. Lines, original trends identified by Walker (1973; solid lines labeled “Walker's upper limit” and “Walker's lower limit”) and theoretically defined cooling-rate limits (dashed lines; from Pinkerton and Wilson, 1994). “Ocean limited” flows reached the ocean and therefore are minimum flow distances (modified from Riker and others, 2009). Red diamonds represent short duration (<12 hours) lava flows erupted from Kīlauea in July (short) and December (long); the latter was confined. Similarly, blue squares represent Mauna Loa eruptions of ~3 weeks duration in 1984 (short) and 1859 (long); the latter were confined.

Effusion rate does exert a fundamental control on the initial advance rates of channelized lava flows (fig. 13*A*; Rowland and Walker, 1990), consistent with treatment of the initial stages of flow emplacement as Newtonian (or Bingham) fluid with constant viscosity (Takagi and Huppert, 2010). This relation is nicely illustrated using data from recent well-observed eruptions, where initial effusion rates for channelized flows have ranged from ~25 to 1,000 m^3/s and corresponding initial rates of flow advance have varied from <0.02 to 3–4 m/s (fig. 13*B*; Kauahikaua and others, 2003). Importantly, both datasets suggest that slope plays only a secondary role in controlling initial rates of flow advance.

These initial flow-advance velocities form a trend that parallels, but is offset toward higher velocities from, the data compiled by Rowland and Walker (1990) on older flows (fig. 13*A*). This discrepancy probably reflects the fact that the observational data for older flows were mainly obtained from distal sites, where flow advance rates are lower because of decreasing flux (from losses along the channel), crust formation (Kerr and Lyman, 2007), and increases in internal lava viscosity because of cooling-induced crystallization. The observed correspondence between flow-advance rate and volumetric flow rate suggests that advance rates inferred from flow features, such as runup heights on tree molds (Moore and Kachadoorian, 1980), superelevations on channel bends (Heslop and others, 1989), clinker size, and lava crystallinity, can be used to estimate effusion rates from older eruptions

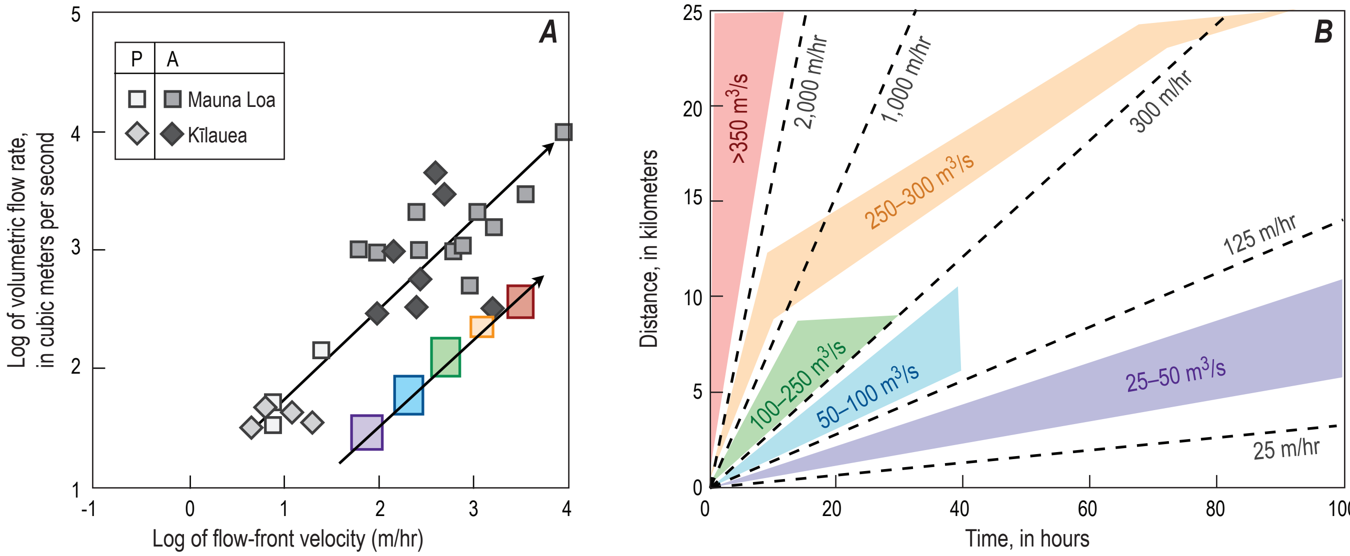

Figure 13. Effusion-rate controls on flow advance rates. *A*, Volumetric flow rate versus flow-front velocity estimated for historical lava flows (diamonds and squares; data from Rowland and Walker, 1990) and constraints from well-observed recent flows (colored boxes; data from Kauahikaua and others, 2003). Offset between two datasets may reflect either underestimates of historical flow-advance rates or overestimates of historical flow-front velocities. P=pāhoehoe, A=ʻaʻā. *B*, Distance versus time for recent lava flows (modified from Kauahikaua and others, 2003). Individual fields are labeled for volumetric flow rates; dashed lines, flow-front velocities. Colors in figures 13*A* and 13*B* refer to same data.

(for example, Soule and others, 2004). These data also highlight the importance for hazard assessment of obtaining accurate estimates of effusion rate during early stages of eruptive activity (see next subsection “Real-Time Flow Monitoring and Hazards Assessment”).

Observational data on lava flow advance also show that the position of eruptive vents relative to local populations affects the relative impact of lava flow eruptions because lava flows typically advance more slowly when farther from their source vents (Kilburn, 1996). For this reason, eruptions from low-elevation flank vents pose much greater hazards than those from vents high on the rift zones of Kīlauea and Mauna Loa, as illustrated by the attempts at barrier construction during the 1955 and 1960 Kīlauea eruptions. Additionally, unusually rapid advance of lava flows was reported for the eruptions of Hualālai volcano in 1801 and Kīlauea volcano in 1823. Both of these lava flows were erupted from vents near the coast (fig. 1); thus, these accounts may derive more from the proximity of the eruptive vents to observers (Kauahikaua and others, 2002; Soule and others, 2004) than from any unusual properties of the erupted lava (Baloga and others, 1995).

Real-Time Flow Monitoring and Hazard Assessment

A compelling reason to study the formation and evolution of lava channels is to construct predictive models of flow paths, advance rates, and areal extents. Existing models of lava flow behavior include empirical relations (for example, Walker, 1973; Pinkerton and Wilson, 1994), probabilistic models of inundation (for example, Kauahikaua and others, 1995, Kau and Truesdell, 1999; Crisci and others 2010), thermal models (Crisp and Baloga, 1990, 1994), rheologic models (Hulme, 1974; Dragoni and Tallarico, 1994; Kilburn, 2004), parametric models (Griffiths and others, 2003; Lyman and others, 2005, Lyman and Kerr, 2006; Cashman and others, 2006; Kerr and others, 2006; Robertson and Kerr, 2012), and (one-dimensional) coupled fluid dynamical and thermal models (Harris and Rowland, 2001). The array of approaches reflects the complexity of the processes involved. Key data required for all models, however, are volumetric lava flux (effusion rate) at the vent and high-resolution topography.

Average lava effusion rates, by definition, can be determined only in hindsight, using measured flow volumes and known eruptive durations (for example, VL 508, 509; Tilling and others, 1987; Wolfe and others, 1988). More useful during eruptive crises, though more difficult to obtain, are daily, or even hourly, measurements of instantaneous effusion rates. Effusion rates can be estimated for flow through individual lava channels if flow velocities and channel geometries are well known (for example, Lipman and Banks, 1987). These estimates, however, are complicated by uncertainties in the bulk density of the erupted lava, by difficulties in estimating channel depths, and by the need to measure different branches of complex channel networks simultaneously. For this reason, interest has been increasing in creating remote sensing tools for effusion rate measurements (reviewed in Harris and others, 2007).

Digital topography (DEMs) forms the basis of all probabilistic models of flow inundation and is required for statistical modeling of, and response to, ongoing eruptions. In Hawaiʻi, GIS analysis of lava sheds, or topographic areas within which lava will be confined for specific vent positions (Kauahikaua and others, 1995, 1998, 2003), have proven effective in both general hazard mapping and guiding emergency-response efforts (fig. 4). Hazard maps for a specific region can be created by an analysis of "paths of steepest descent" within lava sheds (for example, Costa and Macedonio, 2005; Kauahikaua, 2007). Major challenges to developing predictive models using this approach include obtaining the DEM resolution required to anticipate the advance of thin flows and the need for frequent updating of DEMs to account for changes in local topography created by the lava flows themselves (for example, Mattox and others, 1993). From this perspective, new high-resolution DEMs available from repeat lidar surveys provide the optimal resolution (for example, Favalli and others, 2010), but such surveys are generally impractical from the perspective of either cost or time required for data processing (reviewed by Cashman and others, 2013). More cost effective are photogrammetric surveys (for example, James and others, 2010) that provide near-real-time digital elevation data on advancing flow fronts, although these surveys are generally limited in spatial coverage.

From a theoretical perspective, predictive models must incorporate the balance between thermal and dynamical controls on flow advance rate (for example, Griffiths, 2000), as well as considerations of the internal organization of lava flows, particularly in relation to flow focusing and channel formation. Laboratory experiments performed using analogue materials have provided useful insights into this problem. Early experiments by Hulme (1974) used isothermal viscoplastic fluids to explain channel formation; levees were assumed to form where lateral flow was inhibited by the yield stress of the suspension. Lava flows, however, are not isothermal. Experiments that examine the flow of solidifying (cooling) Newtonian fluids through preexisting channels define two regimes: a "tube" regime at low flow rates, and a "mobile crust" regime at high flow rates (Griffiths and others, 2003). The boundary between these regimes is determined by a critical value of the combined parameter $\theta=\psi(R_a/R_0)^{1/3,}$ where $\psi=U_0t_s/H_0$ is the ratio of a surface solidification timescale t_s to a shearing timescale H_0/U_0; H_0 and U_0 are the flow depth and centerline surface velocity in the absence of solidification, respectively; R_a is a Rayleigh number, and R_0 is a constant (Griffiths and others, 2003). This scaling appears applicable to the interpretation of basaltic lava flow emplacement in other terrestrial (Ventura and Vilardo, 2008) and submarine (Soule and others, 2007) settings.

In a mobile-crust (open channel) regime, the crust width d_c is always less than the channel width W but increases as $d_c{\sim}W^{5/3,}$ except where the flow accelerates through constrictions, around bends, or over slope breaks (Cashman and others, 2006). The condition of $d_c=W$ defines the tube regime. Recent experiments have extended this approach to examine solidifying viscoplastic flows (Robertson and Kerr, 2012), which show the same criteria as for tube development but in which a single mobile-crust regime is replaced by a progression from a shear- to a plug-controlled regime with increasing flow rate.

In reality, lava flows rarely occupy predefined channels; thus, channel formation must also be considered. The laboratory studies presented above have given rise to two end-member models of channel and levee formation: (1) channel width controlled by cooling and solidification of Newtonian fluids (Kerr and others, 2006) and (2) channel width controlled by the rheologic properties of an isothermal Bingham fluid (Hulme, 1974; Sparks and others, 1976). In part, these models derive from differences in the initial properties of lava erupted in Hawai'i and at Mount Etna: the dynamics of Hawaiian lava flows, which are erupted at near-liquidus temperatures, appear to be governed largely by initial conditions that control near-vent cooling and crust formation (Griffiths and others, 2003), whereas Etna lavas typically are very crystalline and thus dominated by the constraints of non-Newtonian rheologies (Hulme, 1974). These models have yet to be combined to examine channel formation in solidifying viscoplastic flows.

Pāhoehoe Flows

Macdonald (1953) provided the first systematic description of pāhoehoe flow types. He linked the high vesicularity of most pāhoehoe flows to the high fluidity of the constituent lava. Styles of pāhoehoe flow advance depend on the flow rate and physical properties of the lava (Macdonald, 1953; Gregg and Keszthelyi, 2004). Rapidly advancing pāhoehoe sheet flows move continuously with a rolling motion as the faster upper surface travels over the lower part of the flow. In contrast, slowly moving flow lobes advance erratically, with times of stagnation accompanied by internal inflation. At intermediate flow rates, the growing crust is torn or fractured to produce a new flow lobe.

Scientific advances from almost three decades of flow mapping at Kīlauea include documentation of (1) the importance of flow inflation in pāhoehoe flow fields (Hon and others, 1994), (2) the effectiveness of initial flows in confining and displacing subsequent flow lobes (Mattox and others, 1993), (3) the evolution of lava tube systems (Kauahikaua and others, 2003), (4) the relation of feeder tubes to flow field development (Mangan and others, 1995a; Heliker and others, 1998), and (5) the pressure balance between the summit reservoir and feeder dikes (Kauahikaua and others, 1996). These processes combine to control the morphology of pāhoehoe flow fields.

Formation of Pāhoehoe Flow Fields

The formation of the Kalapana flow field in 1990 stimulated a major advance in understanding pāhoehoe flow emplacement. Here, lava flows advancing into the community of Kalapana were mapped in great spatial and temporal detail (Mattox and others, 1993). The flows were of two dominant types: large primary flows and smaller breakouts from the primary flows. The primary flows advanced rapidly as a sheet and later developed robust lava tubes; as such, they were the main conduit for lava transport into the area. Breakouts from the primary flows formed secondary flows that advanced slowly and developed tubes that were transient; however, the persistence of breakout-fed secondary

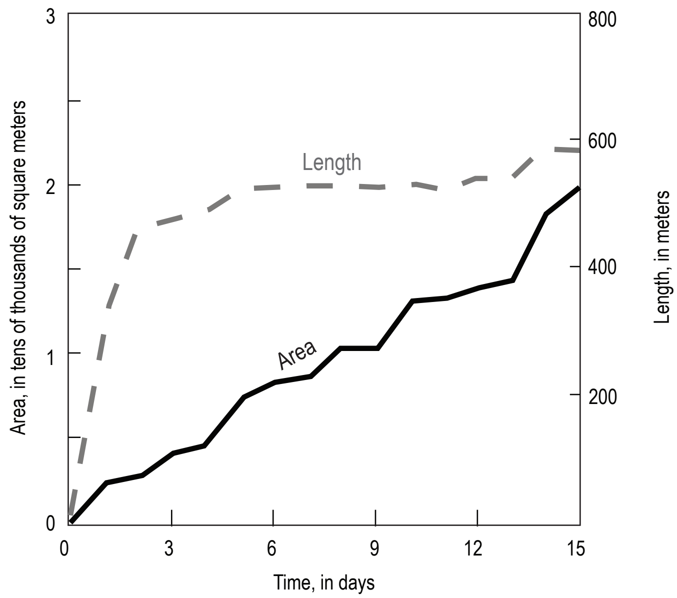

Figure 14. Comparison of variations in length and area versus time for 1990 Kalapana flow. Curve is approximately linear and provides a measure of lava flux (redrafted from Mattox and others, 1993).

flows meant that their volume eventually exceeded that of the primary flows. Additionally, despite erratic advance of individual primary and secondary lava flows, the aerial daily coverage with lava was approximately constant as long as the magma supply remained constant. On the low (2°) slopes where lava flow inflation is the dominant flow emplacement mechanism, planimetric area thus provides a better estimate of effusion rate than does lava-flow length (fig. 14).

Lava-Flow Inflation

The detailed examination of primary flow emplacement by Hon and others (1994) documented the process by which inflated pāhoehoe lava flows form. Lava flows inflate when the transport rate of lava from upslope exceeds the advance rate at the flow front. Because flows advance more slowly on low slopes than on steep slopes, flow inflation is most pronounced on low slopes. Lava flow inflation can occur either within small lobes or across large sheets. Sheet flow inflation causes flows to expand by successive breakouts at both the flow front and around sheet flow margins (for example, Hon and others, 1994; Hoblitt and others, 2012). This pattern of areal expansion explains the relations between areal coverage and time in figure 14.

Actively inflating lava flows have four spatially and rheologically distinct layers: a basal crust, a liquid core, and a crust comprising a lower, partially solidified viscoelastic layer (T=*800–1,070 °C*) and an upper (cooler) brittle layer (fig. 15*A*). If the lava supply remains constant, both the lower and upper crustal layers increase in thickness (H_s) over time *(t)* at a rate controlled by conductive cooling, that is, $H_s=C\sqrt{t}$. Here, C is a constant that Hon and others (1994) took to be 0.0779 for the upper crust, on the basis of cooling measurements in Makaopuhi lava lake. For application to different field situations, this value should be modified depending on both lava vesicularity and local moisture (for example, Keszthelyi, 1995b; Cashman and Kauahikaua, 1997; fig. 15*C*). The lower crustal layer is assumed to grow at 70 percent of the rate of the upper crustal layer because of contact with the underlying rock (Hon and others, 1994). The relation between the thickness of solidified lava and time can be inverted to yield flow duration from the measured thickness of the upper crustal layer of actively inflating flows (Hon and others, 1994) or solidified flows (Cashman and Kauahikaua, 1997). Importantly, the thickness of the liquid core (L) appears to stabilize after about 100 hours and can therefore be measured if the total flow thickness and upper-crustal-layer thickness (H_{su}) are known: $L=H-1.7H_{su}$. Measured liquid cores in the Kalapana flows did not exceed 2 m in thickness and were more commonly about 1m thick (Hon and others, 1994; Cashman and Kauahikaua, 1997). Small sheet flows commonly maintain fluid-core thicknesses of only a few tens of centimeters (fig. 15*B*; Hoblitt and others, 2012).

For a flow to inflate internally, the outer crust must either deform or break. Tensile stresses in the brittle crust cause it to fracture when the flow of lava contained by the crust exceeds the tensile strength of the crust (for example, Hoblitt and others, 2012). Successive fractures at either the sheet flow margins or central axial crack thus record the process of lava flow inflation. Crack formation in inflating lava flows has been equated with fracture propagation in the formation of columnar joints (for example, Aydin and DeGraff, 1988; Grossenbacher and McDuffie, 1995). One common feature is internal banding of inflation cracks at a centimeter scale, a feature that has been attributed to escape of hot gases (Nichols, 1939), lava oozing in from the crack (Walker, 1991), or various failure mechanisms in the brittle and ductile parts of the flow (Hon and others, 1994). A recent study, however, indicates that the banding may instead reflect subtle changes in inflation rate, thereby highlighting the unsteadiness of many lava-emplacement processes (Hoblitt and others, 2012). Unsteady inflation is also illustrated by patterns of vesicle distributions within the upper crust of inflated flows (fig. 15*C*), where zones of small vesicles record times of temporary pressure reduction within in the tube system because of pauses in lava supply (Cashman and Kauahikaua, 1997).

Mapping of structures within inflating pāhoehoe flows has also proven important for understanding active lava-transport systems. Elongate tumuli form over primary lava feeder networks, and hummocky tumuli characterize areas where tubes have slowed or stalled (Hon and others, 1994). These inflation features generate a distinctive surface morphology that has permitted identification of flow inflation in many other terrestrial (Chitwood, 1993, 1994; Keszthelyi and Pieri, 1993; Atkinson and Atkinson, 1995; Stephenson and others, 1998) and submarine (Applegate and Embley, 1992; Gregg and Chadwick, 1996; Chadwick and others, 1999, 2001) basaltic lava flow fields, as well as in some large igneous provinces (for example, Self and others, 1996, 1998; Coffin and others, 2000). Recognition of inflated flow structures in many flood-basalt lava flows has provoked a fundamental re-examination of flood basalt emplacement processes.

Lava Tubes

Lava tubes are integral to the formation of pāhoehoe flow fields and are a signature feature of Hawaiian volcanism. Described by Finch as "one of the most interesting and picturesque of volcanic formations" (VL 82), lava tubes are ubiquitous in pāhoehoe flows of all sizes and may occur in ʻaʻā flows as well (for example, Dutton, 1884). Titus Coan first recognized not only the formation of lava tubes but also their importance in insulating flows and thus permitting long-distance transport of lava (Wright and Takahashi, 1989). Coan's observations persuaded Dana of the importance of lava tubes, despite his original views that lava flows were entirely fissure fed (Dana, 1891). It was also recognized that lava tubes could form by surface crust formation over a lava channel (river) or from lava-filled cracks (VL 82). Early observers commented on the similarities and differences between plan-form-geometry tube-fed lavas and rivers, noting that, like rivers, lava follows topography but that, unlike rivers, lava tubes tend toward distributary systems, such that the discharge diminishes with distance from the vent (VL 82).

Lava-Tube Formation

Modern studies of Hawaiian lava tubes began with the 1969–74 Mauna Ulu eruption, in which tube formation was common and easy to observe (for example, Peterson and Swanson, 1974; Greeley, 1987; Peterson and others, 1994). Observational studies of lava-tube formation during the ongoing Puʻu ʻŌʻō eruption have taken advantage of stable lava tube systems that have extended several kilometers and persisted for many months (for example, Kauahikaua and others, 1998, 2003; Orr, 2011).

Lava tubes are characterized by the presence of a stationary solid crust over fluid lava. In active lava channels, crust formation may initiate at the vent when effusion rates are low, may propagate upchannel from the flow toe when effusion rates are higher, or may initiate in the midsection of a channel at points of channel constrictions or blockages (for example, Greeley, 1987; Peterson and others, 1994; Cashman and others, 2006). Lava tubes on steep slopes typically have headroom (are unfilled) and so are prone to roof collapse. Sites of roof collapse (skylights) permit both sampling and direct measurements of flow temperature, vesicularity,

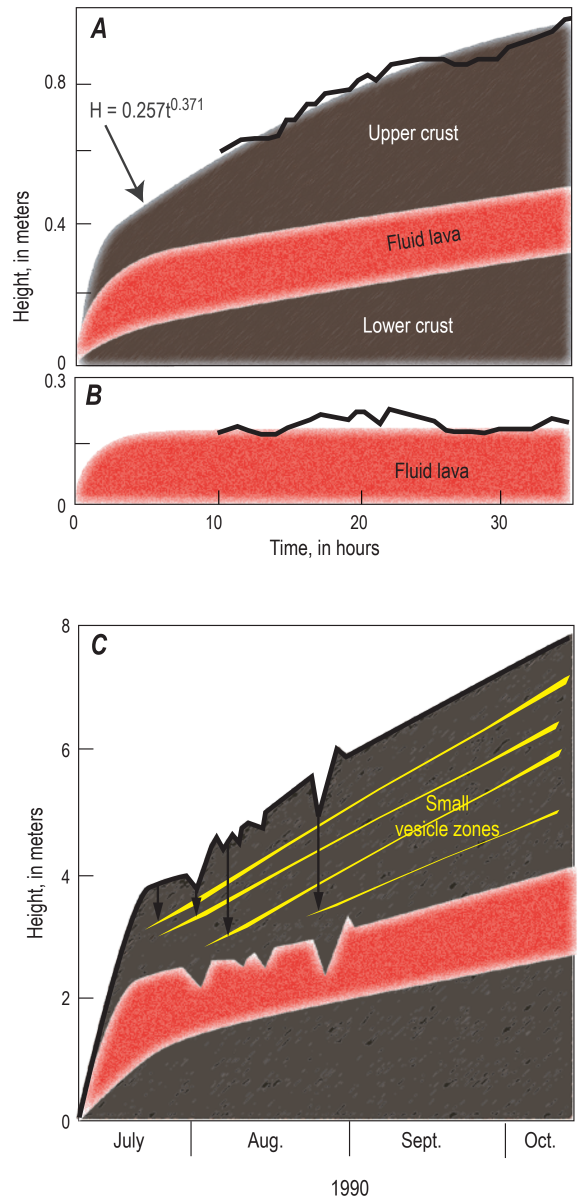

Figure 15. Lava flow inflation. *A*, Measured (dark line) and power-law approximated height of inflated pāhoehoe sheet flow versus time. Extent of fluid lava core (pink) and lower crust were calculated according to constraints of Hon and others (1994). *B*, Inferred fluid lava core versus time (redrafted from Hoblitt and others, 2012). *C*, Long-term flow height of inflated lava flow versus date in 1990. Fluctuations in elevation of upper crust are due to temporary pauses in lava supply, which caused a reduction in internal pressure of lava tube and formation of thin vesicular layers ("small vesicle zones") within solidified crust. General correspondence between vesicular layers and loss of lava tube pressurization suggest that textural variations in upper crust record dynamics of lava flow emplacement (modified from Cashman and Kauahikaua, 1997).

velocity, and (under the right conditions) depth of the lava stream (Cashman and others, 1994; Kauahikaua and others, 1998).

Lava tubes can also form within inflating pāhoehoe flows as sheet flows cool to form enclosing crusts. Inward cooling from the sheet flow surfaces focus the flow into central lava tubes (fig. 15; Hon and others, 1994). The shallow slopes that promote lava flow inflation also inhibit lava tube drainage, so that these tubes tend to remain filled (Kauahikaua and others, 1998, 2003). If flow through the tube is maintained, then continued crustal growth creates either elongate or hummocky tumuli, depending on the rate and steadiness of lava supply. Elongate tumuli mark the sites of steady and persistent lava transport; hummocky tumuli form as lava supply rates wane and (or) interior tube networks start to break down. Ubiquitous lava squeezeups and breakouts from elongate tumuli show that lava within the tubes is moderately overpressurized (Kauahikaua and others, 2003; Hoblitt and others, 2012). In contrast, breakouts from hummocky tumuli are degassed, as illustrated by the formation of dense "blue glassy" lava flows (for example, Hon and others, 1994; Oze and Winter, 2005).

Thermal Efficiency of Lava-Tube Systems

Samples collected along lava tube systems confirm the inferences by early observers that lava tubes are efficient insulators and thus permit the transport of lava over long distances with minimal cooling (for example, Helz and others, 2003), as illustrated by the tube-fed structure of many of the longest terrestrial lava flows (for example, Cashman and others, 1998; Self and others, 1998), and, possibly, many of the longest planetary flows (Keszthelyi, 1995b).

Application of glass geothermometry suggests average cooling rates of 0.6 °C/km along recent Kīlauea lava-tube systems (Helz and others, 2003), which translates to cooling rates of 6×10^{-4} to 18×10^{-4} °C/s for average flow rates of 1–3 m/s (for example, Cashman and others, 1994; Mangan and others, 1995a; Thornber, 2001). Similar distance-referenced cooling rates have been obtained by analysis of quenched samples from older tube-fed pāhoehoe flows from both Kīlauea (Clague and others, 1995, 1999) and Mauna Loa (Riker and others, 2009), although these studies suggest that thermal efficiency may increase with increasing tube length (including the extreme case of flood basalts; for example, Ho and Cashman, 1997). Harris and Rowland (2009) classified lava tube systems as mature if they have thermal efficiencies of ~0.001 °C/s, and immature if they have cooling rates of 0.01 °C/s, similar to the maximum cooling rates estimated for proximal high-temperature channels (in other words, where cooling is predominantly radiative and not buffered by extensive crystallization).

In his analysis of all possible contributions to the heat budget of lava tubes, Keszthelyi (1995b) found latent heat of crystallization to be the single most important heat source; he also concluded that the effects of conduction (in filled tubes), convection (in open tubes), and rainfall are important, whereas radiative cooling from skylights was negligible. More recent FLIR measurements of skylight temperatures on the Kīlauea flow field, however, suggest that radiative heat loss and forced convection over skylights may also contribute to the overall heat budget of lava tubes (Witter and Harris, 2007). Because skylight formation appears to be controlled by both the underlying slope and the stability of lava supply from upslope, steady lava supply and gradual slopes should promote thermally efficient lava transport.

Lava Flux Through Tubes

The flow rate of lava through tube systems can be determined using measurements of very low frequency (VLF) electromagnetic induction, from which the cross-sectional area of dense rock equivalent (DRE) lava can be calculated (Zablocki, 1978; Jackson and others, 1985). Under most conditions, VLF-based lava effusion rates correlate well with rates determined by SO_2 emissions (Sutton and others, 2003). An unusually well constrained example is provided by the VLF measurements of diminishing lava flux from Kaupaianaha that were used to predict the shutdown date of the Kuapaianaha vent (to within a few days; fig. 16; Kauahikaua and others, 1996). Here, the observed linear reduction in volumetric flow rate over time supports a model of shutdown driven by gradual pressure loss within the

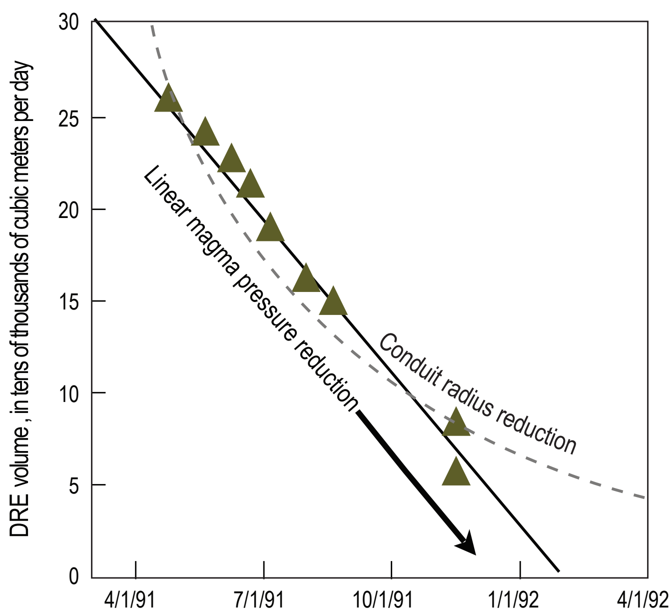

Figure 16. Plot of volumetric flow rate (measured by very low frequency electromagnetic induction) through a lava tube fed from the Kupaianaha vent of Kīlauea Volcano versus time. DRE, dense-rock equivalent. Steady decrease in flow rate was used to predict shutdown of Kupaianaha vent; comparison of conduit-freezing and pressure-reduction models indicates that pressure-reduction model provides a more reasonable fit to data (redrafted from Kauahikaua and others, 1996).

lava tube system, rather than inward solidification of a cylindrical conduit.

An interesting observation is that lava levels within some open tubes (those with headroom) decrease steadily over time, despite maintaining a constant flow width and DRE flux, strongly suggesting erosion of the tube floor. On one occasion, changes in flow depth could be monitored directly, yielding erosion rates of 10 cm/day over several months (Kauahikaua and others, 1998). This rate can be modeled as thermal erosion by assuming steady forced convective heat transfer by laminar channel flow at a large Peclet number (Kerr, 2001, 2009). The predictive capability of a thermal-erosion model does not, however, rule out the possibility that some component of the erosion is mechanical.

Tube-Generated Flow Features

Lava tube systems evolve over time and undergo various types of morphologic alterations. Collapse of lava tube roofs can trigger breakouts that may be violent and characterized by “a sudden outburst of lava, throwing great blocks of cooled lava, with the molten lava, into the air” (VL 520). Persistent weaknesses in tube roofs can produce surface structures such as hornitos, rootless shields, and shatter rings (Kauahikaua and others, 2003). Hornitos form when open tubes fill temporarily with lava and contain sufficient gas to forcibly expel lava clots through a single opening in the tube roof; rootless shields form where perched lava ponds overflow repeatedly over a period of weeks and produce an accumulation of thin (and commonly shelly) pāhoehoe flows; and shatter rings are approximately circular areas of broken pāhoehoe that form over active lava tubes in places characterized by abrupt decreases in flow velocity within the tube (Orr, 2011). Because a decrease in flow velocity must be accompanied by an increase in lava thickness within the tube (by mass conservation), these places are susceptible to overpressurization and repeated rupture of the tube roof. In other parts of the world, similar features on older flows have been described as “collapsed tumuli” (Guest and others, 1984), “unusual craters” (Summerour, 1989), “craters with raised rims” (Greeley and Hyde, 1972), and “lava ponds” (Atkinson and Atkinson, 1995). The spatial arrangement of hornitos, lava shields, and shatter rings all provide important information on former lava transport paths and flow emplacement dynamics (both longevity and steadiness; Kauahikaua and others, 2003; Orr, 2011).

Drained tubes are commonly decorated with lava stalactites—“some like grapes, some like walking sticks, and some like worms” (VL 345)—whose origin has been the subject of speculation since the earliest Western visitors to lava tubes (for example, Barton, 1884; Stearns and Clark, 1930; Wentworth and Macdonald, 1953). Jaggar (VL 345) described the stalactites as “material of the gas-melted glaze” with an outer coating of “magnetic oxide of iron” and interior vesicles lined with crystals of feldspar and augite. Recent investigations of such “soda straw” stalactites confirm that their surfaces are enriched in oxide phases and that the interior walls are enriched in titanium (Baird and others, 1985; Kauahikaua and others, 2003), both features consistent with remelting of tube walls. Long after flow emplacement, another generation of stalactites may form when rainwater gains access and hydrous sulfates precipitate as this water drips from the ceiling to the floor (Finch and Emerson, 1924; Thornber and others, 1999; Porter, 2000).

Lava-Seawater Interactions

Interactions between flowing lava and the ocean are common in Hawai‘i. These interactions may be peaceful or explosive, depending on the flow style and specific conditions of ocean entry. Accounts of explosive lava-water interactions accompanying the ‘a‘ā-producing eruptions of Kīlauea in 1840 and of Mauna Loa in 1868 and 1919 record both the violence of the explosions and the structure of the resulting fragmental cones (summarized by Moore and Ault, 1965). The violence of the lava-water interaction is believed to reflect the large surface area of rough ‘a‘ā flows. This hypothesis can be tested by examining ‘a‘ā flows that do not generate explosions at ocean-entry sites, such as three discrete sites created during the 1950 Mauna Loa eruption (fig. 3). Here, observers noted “a huge column of steam” but no violent explosions or ash generation. Finch and Macdonald (1950) speculated that the absence of explosions reflected both the high temperature and relatively smooth surface of the 1950 lava streams.

Pāhoehoe flows are typically quiescent when entering the ocean, with lava quenched by seawater shattering only when it reaches the surf zone. Tube-fed pāhoehoe flows can, however, exhibit a range of lava-water interaction styles, including the unusually explosive lava-water interactions at Kīlauea between 1992 and 1994 (Mattox and Mangan, 1997). One condition that promotes explosive activity at pāhoehoe ocean entries is the formation of lava deltas with unstable edges that collapse repeatedly, thereby exposing the lava tube to ocean water. Under these conditions, four types of explosions may occur (Mattox and Mangan, 1997): tephra jets, lithic blasts, bubble bursts, and littoral lava fountains (fig. 17). Tephra jets, the most common types, form when lava from the severed tube is exposed to the surf zone. Prolonged tephra-jet activity at a single site can produce a semicircular agglutinated littoral cone. Lithic blasts form during delta collapse, when seawater contacts newly exposed incandescent rock. These two types of lava-water interaction can be explained by open mixing processes at water/melt ratios of ~0.15 (fig. 17*A*). Bubble bursts, possibly the most spectacular (though relatively mild) form of lava-seawater interaction, form when seawater gains rapid entry into a confined lava tube. Bubble bursts produce fluid bombs, limu-o-Pele (thin glass sheets) and Pele’s hair, all of which can accumulate to form circular agglutinated cones. Littoral fountains are rare but may accompany bubble bursts; they form circular spatter cones inland from the shoreline. Both require confined mixing of lava

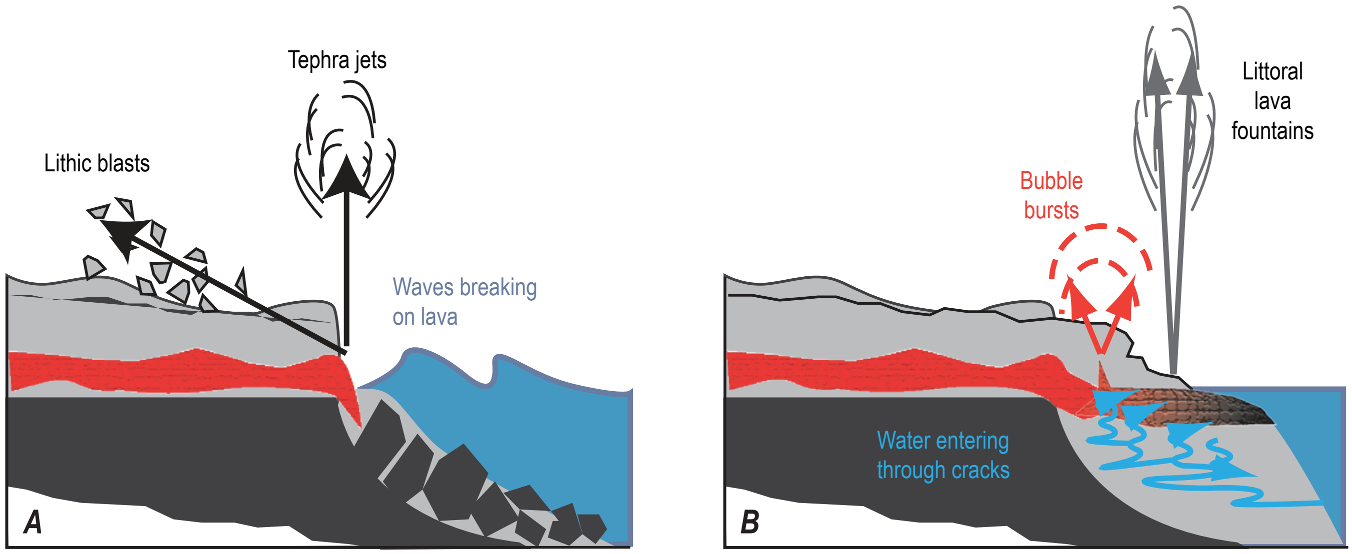

Figure 17. Primary types of lava-water interaction at ocean entries of tube-fed lava flows from Kīlauea Volcano. Lava-water interactions are divided into two primary types. *A*, Open mixing occurs when collapse of lava deltas exposes lava tube to seawater. Waves breaking on exposed lava can produce either tephra jets or, at time of delta collapse, directed lithic blasts. *B*, Confined mixing of seawater and lava occurs when water gains entrance to tube through cracks within lava delta. Under these circumstances, heated steam confined within lava produces spectacular large bubble bursts or, more rarely, energetic littoral lava fountains (redrafted from Mattox and Mangan, 1997).

and water within the tube (fig. 17*B*), with the intensity of the explosion controlled by the rate of water entry, which determines the water/melt ratio. Confined mixing of lava and seawater may explain the large circular littoral cone associated with a prehistoric pāhoehoe flow from Mauna Loa (Jurado-Chichay and others, 1996).

Contributions of Hawaiian Volcanology to Planetary Volcanism

The application of observations of Hawaiian lava flows to the interpretation of planetary lava flow features has a long history. Early geologists (for example, Scrope, 1872) recognized the volcanic origin of lunar surfaces, and Hitchcock (1911) reported that in 1905, W.H. Pickering of Harvard University visited Hawai‘i to make comparisons with lunar features (particularly craters). Since that time, Hawai‘i has served as both a benchmark and a testing ground for developing methods of analyzing planetary surfaces (for example, Greeley, 1974; Carr and Greeley, 1980; Mouginis-Mark and others, 2011; Rowland and others, 2011).

Analysis of surface features to infer the origin of landforms (volcanic geomorphology) is the core of planetary volcanism; the calibrations required for feature interpretation have come from studies of terrestrial basaltic lava flows. Hawaiian lava flows, in particular, have provided an important baseline for mapping the surfaces of the Moon, Mercury, Venus, and Mars; and recent activity in Hawai‘i has provided a context for the interpretation of observed eruptions on Jupiter's moon Io. Hawai‘i has served as a critical testing ground for applying different types of remote sensing to mapping volcanic regions. Hawaiian lava flows are also used to benchmark models of flow emplacement, particularly in order to infer conditions of lava flow emplacement from measured lava flow geomorphologies. Here, we do not attempt a comprehensive review of the ways in which Hawaiian studies have contributed to the study of extraterrestrial volcanism, but instead provide a brief overview of examples where the lava flow features described above have been used to deduce volcanic processes on other planets.

Ground Truthing Remote-Sensing Techniques

Exploration of planetary surfaces has been accomplished almost entirely by remote sensing. For this reason, numerous experiments have been conducted in Hawai‘i to assess the sensitivity of remote-sensing techniques for identifying different types of volcanic features. Flow-surface characteristics can be analyzed by using either multispectral thermal (Kahle and others, 1988) or radar (Gaddis and others, 1989, 1990) data. Thermal data can be used to map individual lava flows or flow lobes but does not work well for larger scale mapping. More useful is radar, which can penetrate clouds (essential for mapping the surface of Venus) and uses an active sensor (which allows data acquisition in permanently shaded regions; see overview by Mouginis-Mark and others, 2011). Radar images highlight topographic features such as faults and caldera walls and can be used to distinguish smooth from rough flow surfaces (Gaddis and others, 1989). In this context, "smooth" and "rough" are defined with reference to the scale of vertical relief (measured as the root mean square of vertical elevation changes) when calibrated for the incidence angle of the radar. New Synthetic Aperature Radar techniques and both airborne laser scanning and terrestrial laser scanning have also been tested at Kīlauea to improve surface-roughness analysis for application to mapping both Mars and Venus (for example, Campbell and Shepard, 1996; Carter and others, 2006; Morris and others, 2008). Development of these remote-sensing techniques has fed back to monitoring of Hawaiian volcanoes, particularly with regard to improved techniques of thermal imaging and analysis (for example, Wright and others, 2010, 2011).

Mapping Planetary Surfaces

Identifying lava flow features on the Moon and inner planets has been a primary goal in mapping planetary surfaces. For example, lunar sinuous rilles first identified by Christian Huygens in 1684, were recognized as bearing a striking resemblance to collapsed lava tubes because of observations during the 1969–74 Mauna Ulu eruption (for example, Greeley 1971; Cruikshank and Wood, 1972; Carr, 1974). Also important has been development of techniques to distinguish ‘a‘ā from pāhoehoe flow surfaces, including not only the surface-roughness techniques described above (Gaddis and others, 1989) but also fractal analysis of lava flow margins (Bruno and others, 1994).

As imaging techniques have improved, so has the quantification of planetary lava-flow features. For example, maps showing the distribution and morphology of lava flows in the Elysium Planitia region of Mars provide information not only on flow aspect ratios but also on their spatial distribution (Mouginis-Mark and Yoshioka, 1998). One important observation from this work is that all the flows erupted <200 km from the summit of Elysium Mons are short (on Martian scales; <70 km), whereas the 11 longest flows have vents >294 km from summit. This observation is similar to eruptive patterns that contribute to the classic so-called “inverted soup bowl” form of basaltic volcanoes in the Galápagos Islands, and may reflect the unusually steep upper flanks (>7°; Wilson and Mouginis-Mark, 2001) of Elysium Mons: flow of lava down steep slopes enhances cooling and crystallization and thus limits flow length (for example, fig. 12).

Increased resolution of Mars surface images provided by the Mars Reconnaissance Orbiter’s High Resolution Imaging Science Experiment permits detailed mapping of meter-scale volcanic features (Keszthelyi and others, 2008). Also possible is analysis of the emplacement conditions of individual lava flows. One example uses tumulus spacing on hummocky pāhoehoe surfaces to infer volumetric flux (Glaze and others, 2005; Bruno and others, 2006). Mars Orbiter Laser Altimeter (MOLA) digital elevation data allow detailed measurements of larger flow features, such as channel and flow dimensions, and changes in flow dimensions as a function of distance from source vents (fig. 18). In an example provided by Glaze and Baloga (2006), both flow width and flow thickness increase as a function of distance from the source vent, consistent with a distributary model of lava transport.

Dynamics of Planetary Volcanism

Some models of terrestrial lava flow behavior have been constructed specifically for application to planetary processes. For example, Hulme’s (1974) widely used model of levee formation in Bingham fluids was constructed specifically with the intent of obtaining both rheologic and dynamic information on planetary lava flows (Hulme, 1982). This Bingham model has been justified because Martian lava flows do not widen at the rate expected for Newtonian fluids (for example, Wilson and Head, 1994). An alternative explanation for the absence of flow widening is levee formation because of marginal flow cooling (Kerr and others, 2006). Consideration of cooling-limited flow widths is important because the cooling histories of planetary flows are not expected to differ significantly from those of terrestrial lava flows (for example, Wilson and Head, 1983, 1994; Head and Wilson, 1986, 1992). More important is the effect of variations in gravity on magma ascent and eruption in planetary and terrestrial environments. Lower gravity will affect both magma ascent and flow across the surface and must be accounted for if observed relations between lava flow geometry and effusion rate are used to estimate the eruptive conditions of planetary flows (for example, Malin, 1980; Crisp and Baloga, 1990; Head and Wilson, 1992). Curiously, Wilson and Head (1994) predicted that cooling-limited flows will be longer by a factor of 6 on Mars than on Earth because the lower gravity means that flows will be thicker for a given yield strength and slope. Alternatively, Rowland and others (2004) suggested that channel-fed flows should be shorter on Mars because lower flow rates will allow more extensive cooling and crystallization per unit distance. In any case, Martian flows, in particular, are generally much larger than current Hawaiian lava flows, more on the scale of flood basalts. Interestingly, however, many flow features seem to scale similarly to those on Earth. One example is the partitioning of lava between stagnant levees and active channels, as determined from analysis of MOLA data (fig. 18) and data from the 1984 Mauna Loa eruption (Lipman and Banks, 1987; fig. 19). Although volumes differ, both flows show increasing lava storage in stagnant levees and overflows with increasing distance from the vent, consistent with observed decreases in volumetric flux along channel networks (fig. 11).

Recent research on Mars outflow channels has also revived discussion of the extent to which planetary surfaces can be sculpted by volcanic (rather than fluvial) processes (for example, Hulme, 1982; Leverington, 2004; Jaeger and others, 2010). Pertinent to this discussion is evidence in Hawai‘i for both thermal (Kauahikaua and others, 1998) and mechanical (Kauahikaua and others, 2002) erosion in lava tubes and channels, consistent with models for the formation of sinuous rilles and analogous features (for example, Fagents and Greeley,

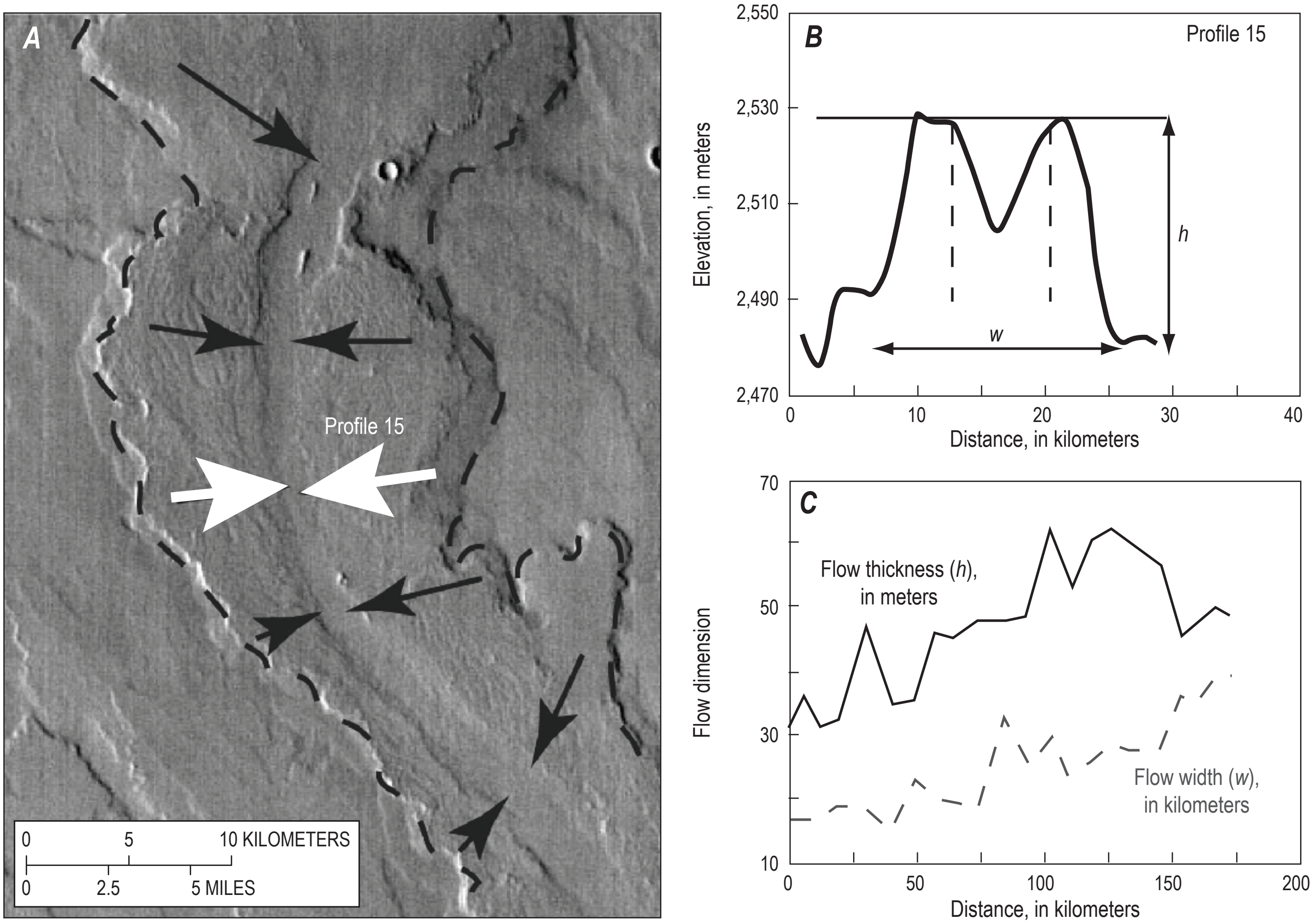

Figure 18. Martian lava flow. *A*, Image of segment of a long channelized lava flow on northern flank of Pavonis Mons (from Baloga and others, 2003). Dashed lines indicate flow margin, black arrows point to central channel, white arrows indicate channel cross section in part *B*. THEMIS infrared image I01739006 from Glaze and Baloga (2006). *B*, Cross section of lava flow, showing marginal levees and central channel (vertical exaggeration, ~400x). *C*, Summary of data along flow showing variations in both flow thickness and flow width as a function of distance from vent. From Baloga and others (2003).

2001). This work has also stimulated mapping of constructional features that may reflect lava-water interaction (for example, Hamilton and others, 2010).

Finally, discovery of active volcanism on Jupiter's moon Io has prompted direct comparison to Hawaiian eruption mechanisms. Repeat observations of volcanic activity on Io allow discrimination of two eruption modes: Promethean and Pillanian (Keszthelyi and others, 2001). Promethean volcanic activity is longlived but pulsatory and produces compound lava-flow fields. This eruptive style, which has been equated directly with that of Puʻu ʻŌʻō during the 1980s, is interpreted to result from shallow magma storage (Davies and others, 2006). Pillanian volcanic activity, in contrast, is characterized by short energetic eruptions that produce large pyroclastic deposits and fissure-fed lava flows; this activity is interpreted to reflect rapid transport of magma from deeper storage regions. An eruption of this type in 1997 involved a 40-km-long fissure and emplacement of two 50-km-long flow lobes (Williams and others, 2001). Estimated effusion rates of 10^3 to 10^4 m^3/s for this eruption exceed those of historical eruptions from Kīlauea but overlap those of the 1950 Mauna Loa eruption. This correspondence, together with the general tendency of eruptions from Mauna Loa to initiate at higher effusion rates than those of Kīlauea (at least historically), raises interesting questions about the general relations between magma-storage depths, volumes, eruption rates, and styles of basaltic volcanism.

Summary

Succinctly summarizing a century of Hawaiian lava flow research is difficult; however, HVO has clearly made a critical contribution to volcanology through careful documentation of the numerous effusive eruptions that have occurred at Mauna Loa and Kīlauea volcanoes since 1900. The value of complete records of flow field emplacement cannot be overstated, because these data provide a baseline for testing hypotheses and models.

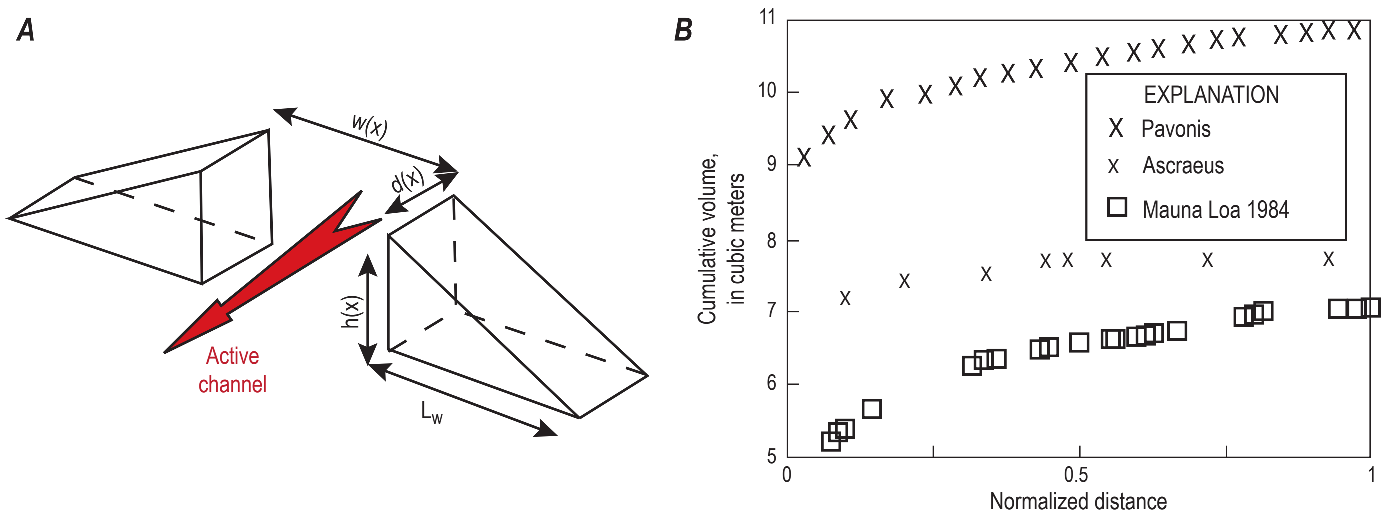

Figure 19. Measurements of lava storage along channels. *A*, Illustration of assumptions used to infer volume stored within marginal levees. *B*, Cumulative stored volume as a function of relative distance along a lava flow. Xs show lava storage along Martian channels (modified from Glaze and Baloga, 2006). Squares show lava storage along the 1984 Mauna Loa channel (calculated from fig. 11*B*). Increases in stored lava volume with distance illustrate extent to which lava transferred through proximal channels is lost to either marginal levee construction or channel overflows.

Eruptions of Mauna Loa in 1984 and of Pu‘u ‘Ō‘ō during 1983 to the present are particularly well characterized, thanks to helicopter support for channel measurements in 1984, and persistent monitoring coupled with expanding technological advances during the past three decades. Detailed maps of these eruptions illustrate the complexity of lava flow fields and provide continued opportunity for analysis and modeling.

Close observations of flowing lava have led naturally to detailed studies of lava properties, particularly the relation between the thermal and rheologic evolution of lava flows and the consequent changes in flow-surface morphology. Early pioneering studies using pyrometers, thermocouples, and simple rheologic models have been supplemented by new field measurement tools, laboratory calibrations, analog experiments, and theoretical analyses. Together, these data highlight the complex physics underlying the basaltic flow-surface morphologies that have fascinated scientists for so long. Observed links between cooling conditions, crystallization, and flow dynamics also underline the importance of linking physical and chemical driving forces in many aspects of volcanology.

Finally, careful observations during lava flow episodes have catalyzed important studies of flow-emplacement dynamics. As a result, the basic characteristics of channelized ‘a‘ā and tube-fed pāhoehoe flows are now well understood. These flow types are distinguished primarily by eruption rate, which controls the relative balance between flow advection and formation of surface crusts.

Channelized flows form when volumetric eruption rates are high. Initial flow advance rates are controlled by effusion rate, although flow advance rates generally decrease with increasing flow distance from eruptive vents because of loss of lava to the channel margins, as well as cooling, crystallization, and consequent increases in lava viscosity. Simple channelized flows reach lengths that are proportional to the volumetric flux. Long-lived flows are not simple, however, because lava channels evolve in space and time as levees form, flows split around topographic obstacles, and lava spillovers create new channel branches. These modifications to channel networks affect both the flow advance rate and the ultimate length achieved by individual flow lobes. Although both probabilistic and single-channel models have been constructed for lava flow prediction, full characterization of channelized lava flows has not yet been achieved.

Tube-fed pāhoehoe flows form when initial volumetric flow rates are sufficiently low that lava crusts form more rapidly than they are disrupted by advection. Because lava is an excellent insulator, tubes can transport lava over long distances with little cooling. Tubes on moderate slopes can erode their base; as a result, lava tubes are open (have headroom). Tubes on low slopes remain filled and therefore cool by conduction through both the upper and basal crust. Breakouts from filled tubes are common and modulate the spatial coverage of pāhoehoe flow fields. The tendency of pāhoehoe flows to inflate from very thin initial to thick final flows requires both very-high-resolution digital terrain models and frequent remapping to anticipate the effect of newly inflated topographic barriers on paths of subsequent lava flows.

A century's perspective is an interesting lens from which to view any field in science. In writing this review, we have been impressed by the prescience of early scientists in identifying topics that have proved to be of fundamental importance not only to volcanology but also to fields such as materials science, fluid mechanics, and planetary science. We have also come to appreciate the wisdom of developing, promoting, and maintaining a natural laboratory for studies of basaltic volcanism and for routine, continuous monitoring of eruptive events. Although the sheer number of scientists who have visited, worked with, and learned from HVO scientists is difficult to quantify, it is through these contacts that HVO has had a global impact on lava flow studies.

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